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Hydrologic Properties and Ground-Water Flow Systems of the Paleozoic Rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, Excluding the San Juan Basin

Regional Aquifer-System Analysis

Professional Paper 1411–B

USGS
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Hydrologic Properties and Ground-Water Flow Systems of the Paleozoic Rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, Excluding the San Juan Basin

By ARTHUR L. GELDON

REGIONAL AQUIFER-SYSTEM ANALYSIS—
UPPER COLORADO RIVER BASIN, EXCLUDING SAN JUAN BASIN

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1411–B

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GALE A. NORTON, ***Secretary***

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Charles G. Groat, *Director*

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) program was started in 1978 after a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA program represents a systematic effort to study a number of the Nation's most important aquifer systems which, in aggregate, underlie much of the country and which represent important components of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system and accordingly transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies, both to develop an understanding of the natural, undisturbed hydrologic system, and of any changes brought about by human activities, as well as to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA program is assigned a single Professional Paper number, and where the volume of interpretive material warrants, separate topical chapters that consider the principal elements of the investigation may be published. The series of RASA interpretive reports begins with Professional Paper 1400, and thereafter will continue in numerical sequence as the interpretive products of subsequent studies become available.

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(Plates are in pocket)

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Multiply	By	To obtain
acre (acre)	0.4047	square hectometer
acre-foot (acre-ft)	0.001233	cubic hectometer
acre-foot per year (acre-ft/yr)	0.001233	cubic hectometer per year
barrel (42 gallons) per day	0.00184	liter per second
barrel per day per foot (bbl/d/ft)	0.00604	liter per second per meter
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per second per mile ($ft^3/s/mi$)	0.01760	cubic meter per second per kilometer
cubic foot per second per square mile ($ft^3/s/mi^2$)	0.01093	cubic meter per second per square kilometer
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
gallon (gal)	3.785	liter
gallon per minute (gal/min)	0.06308	liter per second
inch (in.)	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer
mile per year (mi/yr)	1.609	kilometer per year
pound per square inch (lb/in^2)	0.07037	kilogram per square centimeter
pound per cubic inch (lb/in^3)	0.02768	kilogram per cubic centimeter
square foot (ft^2)	0.0929	square meter
square inch per pound (in^2/lb)	14.2	square centimeter per kilogram
square mile (mi^2)	2.59	square kilometer

Temperature in degrees Fahrenheit (°F) can be converted to temperature in degrees Celsius (°C) by the following equation:

$$°C = 5/9\,(°F - 32)$$

A millidarcy (md) is 0.987×10^{11} centimeter squared.

National Geodetic Vertical Datum of 1929 (NGVD of 1929—A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "Sea Level Datum of 1929."

HYDROLOGIC PROPERTIES AND GROUND-WATER FLOW SYSTEMS OF THE PALEOZOIC ROCKS IN THE UPPER COLORADO RIVER BASIN IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING, EXCLUDING THE SAN JUAN BASIN

By ARTHUR L. GELDON

"All winter long snow falls on its mountain-crested rim, filling the gorges, half burying the forests, and covering the crags and peaks with a mantle woven by the winds from the waves of the sea. When the summer sun comes this snow melts and tumbles down the mountainsides in millions of cascades. A million cascade brooks unite to form a thousand torrent creeks; a thousand torrent creeks unite to form half a hundred rivers beset with cataracts; half a hundred roaring rivers unite to form the Colorado, which rolls, a mad, turbid stream, into the Gulf of California."

John Wesley Powell,
The Exploration of the Colorado River and Its Canyons

ABSTRACT

The hydrologic properties and ground-water flow systems of Paleozoic sedimentary rocks in the Upper Colorado River Basin were investigated under the Regional Aquifer-System Analysis (RASA) program of the U.S. Geological Survey in anticipation of the development of water supplies from bedrock aquifers to fulfill the region's growing water demands. The study area, in parts of Arizona, Colorado, New Mexico, Utah, and Wyoming, covers about 100,000 square miles. It includes parts of four physiographic provinces—the Middle Rocky Mountains, Wyoming Basin, Southern Rocky Mountains, and Colorado Plateaus. A variety of landforms, including mountains, plateaus, mesas, cuestas, plains, badlands, and canyons, are present. Altitudes range from 3,100 to 14,500 feet. Precipitation is distributed orographically and ranges from less than 6 inches per year at lower altitudes to more than 60 inches per year in some mountainous areas. Most of the infrequent precipitation at altitudes of less than 6,000 feet is consumed by evapotranspiration. The Colorado and Green Rivers are the principal streams; the 1964–82 average discharge of the Colorado River where it leaves the Upper Colorado River Basin is 12,170 cubic feet per second (a decrease of 5,680 cubic feet per second since construction of Glen Canyon Dam in 1963).

On the basis of their predominant lithologic and hydrologic properties, the Paleozoic rocks are classified into four aquifers and three confining units. The Flathead aquifer, Gros Ventre confining unit, Bighorn aquifer, Elbert-Parting confining unit, and Madison aquifer (Redwall-Leadville and Darwin-Humbug zones) make up the Four Corners aquifer system. A thick sequence, composed mostly of Mississippian and Pennsylvanian shale, anhydrite, halite, and carbonate rocks—the Four Corners confining unit (Belden-Molas and Paradox–Eagle Valley subunits)—overlies the Four Corners aquifer system in most areas and inhibits vertical ground-water flow between the Four Corners aquifer system and the overlying Canyonlands aquifer. Composed of the uppermost Paleozoic rocks, the Canyonlands aquifer consists, in ascending order, of the Cutler-Maroon, Weber-De Chelly, and Park City–State Bridge zones. The Paleozoic rocks are underlain by a basal confining unit consisting of Precambrian sedimentary, igneous, and metamorphic rocks and overlain throughout most of the Upper Colorado River Basin by the Chinle-Moenkopi confining unit, which consists of Triassic formations composed mostly of shale.

The largest values of porosity, permeability, hydraulic conductivity, transmissivity, and artesian yield are exhibited by the Redwall-Leadville zone of the Madison aquifer and the Weber-De Chelly zone of the Canyonlands aquifer. The former consists almost entirely of Devonian and Mississippian carbonate rocks; the latter consists mostly of Pennsylvanian and Permian quartz sandstone. Unit-averaged porosity in hydrogeologic units composed of Paleozoic rocks ranges from less than 1 to 28 percent. Permeability ranges from less than 0.0001 to 3,460 millidarcies. Unit-averaged hydraulic conductivity ranges from 0.000005 to 200 feet per day. The composite transmissivity of Paleozoic rocks ranges from 0.0005 to 47,000 feet squared per day. Artesian yields to wells and springs (excluding atypical springflows) from these hydrogeologic units range from less than 1 to 10,000 gallons per minute. The permeability and water-supply capabilities of all hydrogeologic units progressively decrease from uplifted areas to structural basins.

Recharge to the Paleozoic rocks is provided by direct infiltration of precipitation, leakage from streams, and ground-water inflows from structurally continuous areas west and north of the Upper Colorado River Basin. The total recharge available from ground-water systems in the basin from direct precipitation and stream leakage is estimated to be 6,600,000 acre-feet per year. However, little of this recharge directly enters the Paleozoic rocks. The recharge from interbasin flow is estimated to be about 1,000 acre-feet per year.

Within the Four Corners aquifer system and Canyonlands aquifer, ground water moves from peripheral and internal highlands to the eastern Great Divide Basin, the confluence of the Yampa and Green Rivers, the San Juan Basin, and the confluence of the Colorado and Little Colorado Rivers (in the Lower Colorado River Basin). Estimated rates of lateral ground-water movement within the Redwall-Leadville zone of the Madison aquifer and the Weber-De Chelly zone of the Canyonlands aquifer range from 0.000001 to 600 feet per day.

Water in the Paleozoic rocks discharges to streams, springs, and wells, is consumed by evapotranspiration where the Paleozoic rocks are at or near land surface, rises into Mesozoic and Tertiary rocks in structural basins along sub-regional flow paths, or flows out of the Upper Colorado River Basin into adjacent hydrologic basins. Total discharge to springs and streams is estimated to equal or exceed 810,000 acre-feet per year. Total withdrawals of water from water, oil, and gas wells is less than 82,000 acre-feet per year. Outflow to the Lower Colorado River Basin, as determined from the combined discharge of springs issuing from Paleozoic rocks in Marble Canyon and the canyon of the Little Colorado River, is less than 170,000 acre-feet per year. Outflows to the Hanna and San Juan Basins, losses to evapotranspiration, and leakage to Mesozoic and Tertiary rocks cannot be quantified at present.

INTRODUCTION

In anticipation of increased water use for the development of coal and oil resources and to meet the needs of an expanding population, the U.S. Geological Survey (USGS) from 1981 to 1990 undertook a systematic appraisal of the ground-water resources of the Upper Colorado River Basin (UCRB). This study, known as the Upper Colorado River Basin Regional Aquifer-System Analysis (UCRB-RASA), was part of a nationwide investigation of regional aquifers by the USGS (Sun, 1986). According to Taylor and others (1983, 1986), specific objectives of the UCRB-RASA were:

1. Identification of aquifers and confining units among consolidated sedimentary rocks of Cambrian to Tertiary age;
2. Determination of the extent, thickness, and hydrologic characteristics of aquifers and confining units;
3. Determination of the water-supply potential of aquifers;
4. Determination of the geochemical characteristics of the ground water;
5. Analyses of the regional ground-water flow systems under steady-state conditions; and
6. Analysis of the flow-system responses to hypothetical ground-water withdrawal or injection.

The results of the UCRB study are presented in USGS Professional Paper 1411, which consists of several chapters. This report, Professional Paper 1411–B, briefly describes the geology of the Paleozoic rocks in the UCRB and extensively describes the hydrologic properties and ground-water flow systems of these rocks. This report synthesizes geological, geochemical, geophysical, and hydrological information from numerous published reports and files of unpublished data. Some original geochemical and aquifer-test data were incorporated into the study to support interpretations. Sources for geologic material used in this report include unpublished petroleum industry borehole logs prepared by the American Stratigraphic Company, lithologic logs provided by the Bureau of Reclamation (written commun., 1983–85), and measured stratigraphic sections contained in published reports cited herein. Sources for hydrologic data used in this report include unpublished drill-stem test, geophysical, and core data compiled by Petroleum Information Corporation, unpublished injection-test information supplied by the Bureau of Reclamation (written commun., 1983–85), unpublished pumping and bailing test information compiled by the Colorado Division of Water Resources, State Engineer's Office, unpublished well and spring data provided by the City of Ouray (written commun., 1988), unpublished material in files of the USGS, and published reports cited herein.

Within this report, hydrologic properties are discussed for the Paleozoic rocks in general and by hydrogeologic unit. A hydrogeologic unit for the purposes of this study is defined as a group of geologic formations and parts of formations that are related stratigraphically and share similar or related lithologic and hydrologic properties. As indicated in table 1, the Paleozoic rocks in this study were classified into 11 hydrogeologic units—four aquifers, three confining units, and subdivisions of aquifers and confining units. This classification was based on the predominant characteristics of each hydrogeologic unit, because the lithology of most units varies extensively throughout the UCRB. An aquifer for the purposes of this study is defined as a group of formations and parts of formations that transmit water to wells and springs throughout most of the UCRB. Confining units consist of a group of formations and parts of formations composed of heterogeneous rock types that generally tend to inhibit vertical ground-water movement but locally may transmit water.

The classification of hydrogeologic units used in this report differs somewhat from previous reports on the hydrogeology of the Paleozoic rocks in the UCRB that were published during the UCRB-RASA (table 1). The current classification scheme represents: (1) An evolution of thought based on increasing amounts of hydrologic and geologic information that became available as the RASA investigation progressed; and (2) an attempt to accommodate guidelines for naming aquifers and confining units in regional-scale studies of complex ground-water systems that were adopted by the USGS several years after the UCRB-RASA began (Laney and Davidson, 1986).

LOCATION OF STUDY

The UCRB, as defined by the Colorado River Compact of 1922, encompasses the drainages of the Green and Colorado Rivers above the mouth of the Paria River at Lees Ferry, Ariz., and the internally drained Great Divide Basin of Wyoming. The UCRB includes about 113,500 mi^2 of land in western Colorado, eastern Utah, southwestern Wyoming, northeastern Arizona, and northwestern New Mexico (pl. 1). Because the San Juan Basin, an area of about 14,600 mi^2, was excluded for separate investigation, this report does not include most of the land in New Mexico and some of the land in southwestern Colorado

TABLE 1.—*Hydrogeologic nomenclature for Precambrian, Paleozoic, and Mesozoic rocks in the Upper Colorado River Basin used in Regional Aquifer-System Analysis reports*

Component geologic units	Hydrogeologic unit of Taylor and others (1986)	Hydrogeologic unit of Lindner-Lunsford and others (1985)	Hydrogeologic unit of Weiss (1990)	Hydrostratigraphic unit of Geldon (1986, 1989a, b, c, d)		Nomenclature in this paper: Hydrogeologic unit	Nomenclature in this paper: Ground-water system
Chinle Formation, Dolores Formation, Moenkopi Formation, Woodside Shale, and Dinwoody Formation; upper State Bridge and Goose Egg Formations	Lower Mesozoic confining layers	Lower Mesozoic confining layers	Not considered	Triassic confining layer		Confining unit consisting of Mesozoic rocks (Chinle-Moenkopi confining unit)	Not applicable
Kaibab, Toroweap, Park City, and Phosphoria Formations; lower State Bridge and Goose Egg Formations	Upper Paleozoic aquifers and confining layers	Upper Paleozoic aquifers	Sandstone and red bed aquifer	Permian shale and carbonate rocks hydrostratigraphic unit	Canyonlands aquifer	Park City–State Bridge zone	
Tensleep, Weber-De Chelly, and Coconino Sandstones; White Rim Sandstone (of Baars, 1962); Wells Formation; Fryingpan and Schoolhouse Members of Maroon Formation (Johnson, 1989; Johnson and others, 1990)				Pennsylvanian and Permian sandstone hydrostratigraphic unit		Weber-De Chelly zone	
Ranchester Limestone Member of Amsden Formation; upper member of Hermosa Formation; Gothic Formation (of Langenheim, 1952); Minturn, Morgan, and Rico Formations; main body of Maroon Formation, undifferentiated Cutler and Supai Formations; Halgaito Shale, Elephant Canyon Formation, Cedar Mesa Sandstone, and Organ Rock Shale of Cutler Group (Baars, 1962); Wescogame Formation and Esplanade Sandstone of Supai Group (McKee, 1982); Hermit Shale				Pennsylvanian and Permian red beds and carbonate rocks hydrostratigraphic unit		Cutler-Maroon zone	
Moffat Trail Limestone Member of Amsden Formation, Paradox Member of Hermosa Formation, Eagle Valley Evaporite, Round Valley Limestone, Manakacha Formation of Supai Group		Upper Paleozoic confining layers	Not considered	Mississippian and Pennsylvanian carbonate rocks and evaporites hydrostratigraphic unit[1]	Four Corners confining unit	Paradox–Eagle Valley subunit	
Horseshoe Shale Member of Amsden Formation, Doughnut Shale, Surprise Canyon Formation, Watahomigi Formation of Supai Group, Molas Formation, lower member of Hermosa Formation, Belden Formation				Mississippian and Pennsylvanian shale and carbonate rocks hydrostratigraphic unit		Belden-Molas subunit	
Humbug Formation, Bull Ridge Member of Madison Limestone, Darwin Sandstone Member of Amsden Formation, upper Mission Canyon Limestone	Middle Paleozoic aquifers	Middle Paleozoic aquifers	Limestone and dolomite aquifer	Mississippian carbonate and clastic rocks hydrostratigraphic unit	Madison aquifer	Darwin-Humbug zone	Four Corners aquifer system
Dyer Dolomite, Ouray Limestone, Gilman Sandstone, Leadville Limestone, main bodies of Madison and Lodgepole Limestones, lower Mission Canyon Limestone, Redwall Limestone				Denvonian and Mississippian carbonate rocks hydrostratigraphic unit		Redwall-Leadville zone	
Elbert, Parting, Darby, and Temple Butte Formations, Cottonwood Canyon Member of Lodgepole and Madison Limestones				Devonian carbonate and clastic rocks hydrostratigraphic unit		Elbert-Parting confining unit	
Mauv Limestone and equivalents, Maxfield Limestone, Lynch Dolomite Gallatin Limestone, Peerless and Dotsero Formations, Manitou Dolomite, Harding Sandstone, Fremont Limestone, Bighorn Dolomite, Ophir Shale, Bright Angel Shale and equivalents, Gros Ventre Formation, upper Lodore Formation	Lower Paleozoic aquifers and confining layers	Lower Paleozoic aquifers and confining layers	Not considered	Cambrian and Ordovician carbonate rocks hydrostratigraphic unit		Bighorn aquifer	
				Cambrian shale hydrostratigraphic unit		Gros Ventre confining unit	
Tintic Quartzite, Tapeats Sandstone and equivalents, Flathead Sandstone, lower Paleozoic Lodore Formation, Sawatch Quartzite, and Ignacio Quartzite	Basal Paleozoic aquifer	Basal Paleozoic aquifer		Cambrian sandstone hydrostratigraphic unit		Flathead aquifer	
Red Creek Quartzite, granitic and metamorphic rocks; Uncompahgre Formation, Uinta Mountain Group, Unkar Group, Nankoweap Formation, and Chuar Group	Not considered	Not considered		Precambrian confining layer		Basal confining unit	Not considered

[1]In Geldon (1986 and 1989c), this unit is referred to as the Pennsylvanian carbonate rocks and evaporites hydrostratigraphic unit because in the area of these two reports, northwestern Colorado, no Mississippian rocks are included in the hydrogeologic unit.

and northeastern Arizona that are in the UCRB. About 100,000 mi^2 of the UCRB are covered by this report. Because hydrogeologic interpretations in the report are based in part on data from areas peripheral to the UCRB, discussions in the report extend beyond the boundaries of the UCRB.

The UCRB, excluding the San Juan Basin, extends from latitude 35°46'N to latitude 43°27'N and from longitude latitude 105×°38'W to longitude 112°19'W. The UCRB, excluding the San Juan Basin, has maximum dimensions of about 560 mi from north to south and about 320 mi from east to west.

The UCRB is sparsely inhabited. According to the U.S. Water Resources Council (1978, p. 5), the total population in the UCRB (including the San Juan Basin) in the mid-1970's was about 344,000. About 68 percent of this population was in western Colorado; 12 percent was in Wyoming; 15 percent was in the Unita Basin area of northeastern Utah; and the remaining 5 percent was in the Canyonlands area of southeastern Utah (see pl. 1 for locations of areas mentioned). The average population density is about four people per square mile, but much of the land is uninhabited. Most of the area's population is clustered in the towns of Grand Junction, Montrose, Delta, Paonia, Glenwood Springs, Aspen, Craig, Steamboat Springs, Meeker, Gunnison, Cortez, and Durango, Colo.; Price, Vernal, Manila, Moab, Monticello, and Blanding, Utah; Rock Springs, Green River, Pinedale, and Kemmerer, Wyo.; Page and Kayenta, Ariz.; and Shiprock and Farmington, N. Mex.

PREVIOUS INVESTIGATIONS

Several comprehensive investigations of water resources in the UCRB preceded the present (RASA) study. Reports by La Rue (1916), Iorns and others (1965), and Liebermann and others (1988) were devoted mainly to discussion of surface water. Surface- and ground-water data compiled for the study by Iorns and others (1965) were published in a separate report by Iorns and others (1964). Additional ground-water data were compiled in a report by Price and Waddell (1973). Price and Arnow (1974) discussed the general availability and quality of ground water in the basin, with emphasis on developing ground-water resources and the impacts from such development. The U.S. Water Resources Council (1978) compiled demographic, physiographic, climatologic, mineralogic, and water-resource data and analyzed water-resource use. Rouse (1967) compiled data on saline springs and other saline inflows to streams in the UCRB. Warner and others (1985) systematically determined ground-water contributions to the salinity of streams in the UCRB.

Many local ground-water investigations in the UCRB provided information used in this study. Areas covered by interpretive reports representative of the ground-water literature for the UCRB are shown in figure 1. Other reports are cited throughout the text and in the "Selected References" section. Some of the information used in this study was obtained from reports on the hydrogeology of areas adjacent to or near the UCRB. In this category were reports by Metzger (1961), Akers and others (1962), Twenter and Metzger (1963), Akers (1964), Bredehoeft (1964), Whitcomb and Lowry (1968), Cooley (1976, 1985), Konikow (1976), Huntoon (1983), Thayer (1983), Downey (1984), Kreitler and others (1985), and McCulley (1985).

Several reports produced during the UCRB-RASA contain hydrologic information about the Paleozoic rocks. Taylor and others (1986) discussed the hydrogeologic framework of the UCRB. Lindner-Lunsford and others (1985) presented a generalized interpretation of the hydrogeology of Paleozoic aquifers in the UCRB. Geldon (1986, 1989c) characterized the hydrogeology of the Paleozoic rocks in northwestern Colorado. Geldon (1989a, 1989b) presented a summary of information contained in this report. Some of the drill-stem test data used in this study were obtained from a report by Teller and Chafin (1986). Most of the hydrologic data used in the study were published in a report by Geldon (1989d). Weiss (1990) presented results of modeling studies of the Paleozoic rocks in the southern part of the UCRB.

SYSTEM OF NUMBERING WELLS AND SPRINGS

Wells and springs are numbered in this report according to the Bureau of Land Management system. The first one or two letters in the site-identification number represent the principal survey meridian. The UCRB is referenced to seven different meridians. Symbols adopted for these meridians in this report are:

G - Gila and Salt River
N - Navajo
NM - New Mexico
S - Sixth
SL - Salt Lake
U - Uintah
UT - Ute

Letters and numbers following the symbol for the principal survey meridian in the site-identification number refer, in order, to quadrant, township, range, section, quarter section, quarter-quarter section, quarter-quarter-quarter section, and number of well or spring within the smallest physical boundary (multiple ground-water sites within the smallest physical boundary are numbered consecutively). Quadrants and divisions of sections are labeled from A to D in a counterclockwise direction starting with the northeast quadrant or section division. Quadrant designations are upper case; section division designations are lower case. Dashes are used to separate township, range, and section designations. As an example, a well numbered SC06–89–09bda_1 is the first well in the northeast quarter of the southeast quarter of the northwest quarter of Section 9, Township 6 South, Range 89 West, in the southwest quadrant of the Sixth principal survey meridian.

FIGURE 1.—Areas covered by ground-water reports in the Upper Colorado River Basin. (Many other interpretive and data reports provided ground-water information for the study area. Reports shown here are either the most comprehensive for the areas they cover or are representative of other reports covering the same area, as of 1987.)

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METHODS OF ANALYSIS

Studying the hydrogeology of the UCRB required extensive searches for sparsely distributed data and diverse approaches to data analysis. Methods of analysis used to describe hydrologic properties are described thoroughly in this section.

POROSITY

Porosity values were obtained by laboratory methods and by interpretation of geophysical logs. In the laboratory, porosity is determined by saturating, weighing, oven drying, and reweighing a sample and then converting the weight of the water lost in drying the sample to the total sample volume (Freeze and Cherry, 1979, p. 337). Most of the 7,665 laboratory-determined values of porosity used in this study were determined from plugs of core taken from boreholes. These values were obtained from unpublished data files of Petroleum Information Corporation. Six of the porosity values were determined from outcrop samples; this information was reported by Hood (1976) and Hood and Patterson (1984).

Porosity also can be determined from sonic, bulk-density, or neutron logs. Sonic logs record the transit time of an acoustic pulse traveling from a probe through the formation and back to a receiver. Bulk-density logs record the intensity of induced gamma radiation backscattered and absorbed by a formation. Neutron logs record the intensity of neutron radiation deflected by a formation. Logging principles and procedures are described by Schlumberger Limited (1972, 1974), Schlumberger Well Services (1984), and Keys (1988). Fox and others (1975) used sonic logs calibrated against a curve of interval transit time versus measured porosity in boreholes in the Bighorn, Wind River, and greater Green River Basins. Data for 11 boreholes in the greater Green River Basin were integrated into this study. As part of the RASA study, Gregory Wetherbee and William Van Liew (U.S. Geological Survey, written commun., 1986) used sonic, bulk-density, and neutron logs, in combination, to determine porosity in 99 borehole intervals; required physical parameters of the sandstone and carbonate rocks studied were based on calibration of geophysical logs against measured core-porosity values from about 20 boreholes.

To determine the regional distribution of porosity within a hydrogeologic unit, unit-averaged porosity values at grid centers on a 10-mi × 10-mi grid or in individual boreholes were contoured. Unit-averaged porosity values were calculated from the median porosity of each rock type present at the grid center or in the borehole and the proportion of the total thickness represented by each rock type. In contouring the unit-averaged values, sparsely distributed measurements of geophysically determined or laboratory-determined porosity in borehole intervals were used as a guide.

No information was available to determine whether reported values of porosity used in this study were total or effective. Considering the methods used to obtain these values, all values were assumed to be effective porosity.

PERMEABILITY

Permeability and hydrologic properties that can be calculated from permeability, such as hydraulic conductivity, are scale-dependent (Dagan, 1986). Values of permeability or hydraulic conductivity differing by orders of magnitude commonly are obtained depending on whether laboratory tests, field tests, or numerical modeling methods are used. For example, Bredehoeft and others (1983) found that field (slug test) and laboratory measurements of the hydraulic conductivity of Cretaceous shale in South Dakota could be one to three orders of magnitude smaller than values of hydraulic conductivity indicated by numerical modeling. At Gibson Dome, in southeastern Utah, regional finite-difference modeling indicated that a horizontal hydraulic conductivity of 0.03 ft/d probably is typical of rocks in the area (INTERA Environmental Consultants, Inc., 1984, p. 82). However, permeameter tests of core samples from a borehole at Gibson Dome indicated horizontal hydraulic-conductivity values ranging from 0.000068 to 0.0042 ft/d, and drill-stem tests of the lower Paleozoic rocks at Gibson Dome indicated horizontal hydraulic-conductivity values ranging from 0.0034 to 0.068 ft/d (Thackston and others, 1984, p. 67).

The problem of the test volume in determining hydrologic properties was addressed by Dagan (1986) from a statistical viewpoint. Dagan considered three scales: the pore scale, the local scale, and the regional scale. The pore scale, which can be assessed by laboratory measurements of core, is on the order of 0.3–3 ft. The local scale, which can be addressed by single-well or multiple-well aquifer tests, is on the order of the aquifer thickness in the horizontal and vertical planes (typically 30–300 ft). The regional scale (measured in miles in the horizontal plane) is much larger than the aquifer thickness and presumably would be conducive to resolution by numerical modeling.

Increasing values of permeability and hydraulic conductivity with increasing scale are best understood in terms of the degree of heterogeneity introduced as the volume of rock under consideration expands. At the pore scale, permeability and hydraulic conductivity depend mainly on the void spaces (pores) between rock grains and, in less than perfect samples, cracks and vugs. At the local scale, permeability measurements take into account large openings, such as joints, fault planes, and solution channels. At the regional scale, lateral variations in lithology also can affect results.

In discussing permeability throughout this report, efforts were made to avoid mixing pore-scale and local-scale values (regional-scale values generally were not available) in the same context. For example, in preparing maps or plots of permeability distribution for a hydrogeologic unit, pore-scale values first were converted to local-scale values, or vice versa, depending on the purpose of the illustration. (An analogous situation would be to prepare a potentiometric-surface map using only heads from the same interval within a lithologically heterogeneous aquifer). Although the reason for the scale dependence of permeability may not have been ascertainable for all data used in this report, following Dagan (1986), laboratory-determined permeability and equivalent laboratory-determined permeability (permeability estimated from porosity, for example) in this report are called pore-scale permeability. Field-determined permeability and equivalent field-determined permeability (permeability estimated from laboratory-determined permeability or hydraulic conductivity) are called local-scale permeability.

Pore-scale permeability usually is determined with a permeameter, a device that measures the rate at which water or gas (typically, nitrogen or air) moves through a sample of rock (Freeze and Cherry, 1979, p. 335–337). Assumptions inherent in the methods used to determine pore-scale permeability and other hydrologic properties studied during the UCRB-RASA are summarized in table 2. For this study, 7,659 measurements of pore-scale permeability based on plugs of borehole core (compiled by Petroleum Information Corp., unpub. data) and 6 measurements based on outcrop samples (reported by Hood, 1976, and Hood and Patterson, 1984) were available.

TABLE 2.—*Types of aquifer tests and methods of analysis used in determining hydrologic properties of Paleozoic rocks in the Upper Colorado River Basin*

[T, transmissivity; S, storativity; K, hydraulic conductivity; k, intrinsic permeability; N/A, not applicable]

Category	Type of test	Assumed test conditions[1]			Test phase	Where data obtained	Methods of analysis			Hydrologic properties
		Constant	Changing	Confining layer			Type-curve matching	Straight-line	Direct solution of an equation	
I	Pumping well	Discharge	Head	Nonleaky	Production	Production well	Theis (1935)	Cooper and Jacob (1946)	Lohman (1979)	T
	Pumping well	Discharge	Head	Nonleaky	Recovery	Production well	Theis (1935)	Theis (1935), Cooper and Jacob (1946)	N/A	T
	Pumping well	Discharge	Head	Nonleaky	Production, recovery	Observation well	Theis (1935)	Cooper and Jacob (1946)	N/A	T, S
	Pumping well	Discharge	Head	Leaky	Production	Production well	Hantush (1960)	N/A	N/A	T
	Pumping well	Discharge	Head	Leaky	Production, recovery	Observation well	Hantush (1960)	N/A	N/A	T, S
II	Flowing well	Head	Discharge	Nonleaky	Production, recovery	Production well	Jacob and Lohman (1952)	Jacob and Lohman (1952)	N/A	T
	Flowing well	Head	Discharge	Leaky	Production	Production well	Hantush (1959)	N/A	N/A	T
	Drill stem	Discharge	Head	Nonleaky	Recovery	Production well	N/A	Horner (1951)	Earlougher (1977); Geldon (1989c)	k
III	Pressure injection	Head, discharge	Not relevant	Not relevant	Entire	Injection well	N/A	N/A	Bureau of Reclamation (1974)	K
	Slug injection	Fluid volume	Head	Not relevant	Entire	Injection well	Cooper and others (1967)	Ferris and Knowles (1963)	N/A	T
IV	Permeameter	Fluid volume	Head	Not relevant	Entire	Rock samples	N/A	N/A	Freeze and Cherry (1979)	k
	Permeameter	Discharge, head	Not relevant	Not relevant	Entire	Rock samples	N/A	N/A	Freeze and Cherry (1979)	k

[1]All analytical methods for categories I and II assume infinitesimal well diameter and instantaneous release from storage.

Additional pore-scale permeability values were obtained from equations relating permeameter-determined permeability to porosity, as was done by Bredehoeft (1964) in a study of porosity and permeability in the Tensleep Sandstone in the Bighorn Basin, Wyo. (location shown on pl. 1). Porosity was plotted against permeameter-determined permeability by hydrogeologic unit and rock type for 7,516 samples. This resulted in the development of nine equations for estimating pore-scale permeability from porosity (table 3). These nine equations update versions appearing in an article by Geldon (1985b), based on an improved understanding of the stratigraphy and a more accurate identification of the sampled intervals. Regardless of hydrogeologic unit, the results of this study indicate that there generally is vague to excellent correlation between porosity and pore-scale permeability in sandstone, but no correlation to little correlation between these properties in carbonate rocks. As discussed later in this report, this observation is interpreted to indicate that pore-scale permeability in sandstone is related mainly to grain size, sorting, and degree of cementation, whereas pore-scale permeability in carbonate rocks depends mostly on fractures and vugs.

Drill-stem-test results from reports by Woodward-Clyde Consultants (1982) and Teller and Chafin (1986) and from unpublished files compiled by Petroleum Information Corporation furnished 405 values of local-scale permeability. Most of these values (258) were determined by the method of Horner (1951). As described by Bredehoeft (1965), this method involves plotting pressure in the well bore during a shut-in period against log $[(t + \Delta t)/\Delta t]$. Permeability is calculated from the straight-line portion of the plot with the following equation:

$$k = \frac{162.6\ qv\ \log[(t+\Delta t)/\Delta t]}{h(p_o - p_w)} \quad (1)$$

where

k = permeability, in millidarcies;

q = discharge rate, in barrels/day;

p_w = well-bore pressure, in pounds per square inch;

p_o = undisturbed formation pressure, in pounds per square inch;

h = thickness of the tested interval, in feet;

v = viscosity, in centipoise.

t = flow period, in minutes; and

Δt = shut-in period, in minutes.

All of the variables necessary to solve equation 1 for permeability, except viscosity and discharge, generally are measured directly. Viscosity usually is estimated from the temperature of

TABLE 3.—*Equations relating porosity to pore-scale permeability for hydrogeologic units composed of Paleozoic rocks in the Upper Colorado Basin*

[<, less than; p, porosity]

	Hydrogeologic unit	Rock type	Equation[1]	Correlation coefficient (R)	Number of observations
Canyonlands aquifer	Park City–State Bridge zone	Sandstone	$k = 0.022\ e^{0.41p}$	0.67	30
		Dolomite	$k = 0.0091\ e^{0.46p}$	.83	521
		Limestone		<.60	76
	Weber-De Chelly zone	Sandstone	$k = 0.017\ e^{0.46p}$	.87	2,636
	Cutler-Maroon zone	Sandstone	$k = 0.017\ e^{0.35p}$	.66	292
		Limestone		<.60	195
Four Corners confining unit	Paradox–Eagle Valley subunit	Sandstone	$k = 0.035\ e^{0.25p}$	.61	37
		Dolomite		<.60	722
		Limestone	$k = 0.019\ e^{0.37p}$	.62	1,649
		Shale		<.60	41
		Anhydrite		<.60	15
	Belden-Molas subunit	Dolomite	$k = 0.017\ e^{1.2p}$	.95	7
		Limestone		<.60	10
Madison aquifer	Darwin-Humbug zone	Sandstone		<.60	19
		Dolomite		<.60	9
	Redwall-Leadville zone	Dolomite		<.60	435
		Limestone		<.60	513
	Elbert-Parting confining unit	Sandstone	$k = 0.010\ e^{0.69p}$	.77	125
		Dolomite		<.60	89
	Bighorn aquifer	Dolomite	$k = 0.0032\ e^{0.74p}$	.81	45

[1]No equation shown if the regression coefficient was less than 0.60.

the water in the test interval. During this investigation, viscosity usually was estimated from equations given by Teller and Chafin (1986, p. 9). These equations are:

$$\mu = 1.93 - [0.818 \times \mathrm{Log}(0.556 \times T_B - 22.8)], \quad T_B = 50° \text{ to } 120° \tag{2}$$

$$\mu = 0.935 - [0.367 \times \mathrm{Log}(0.556 \times T_B - 57.8)], \quad T_B = 120° \text{ to } 425° \tag{3}$$

where

μ = dynamic viscosity, in centipoise; and

T_B = temperature in the test interval, in degrees Fahrenheit.

If the temperature in the test interval was not measured, it was estimated from the thermal gradient in the borehole and the test-interval depth. If the thermal gradient was unknown, it was estimated from figure 2, a map of thermal gradients prepared from temperatures recorded during drill-stem tests (Woodard-Clyde Consultants, 1982; Teller and Chafin, 1986; Petroleum Information Corporation, unpublished) and from published thermal-gradient information in reports by American Association of Petroleum Geologists and U.S. Geological Survey (1976), Barrett and Pearl (1977), Law and others (1980), MacMillan (1980), Law and Smith (1983), Bostick and Freeman (1984), and Nuccio and Johnson (1984).

Drill-stem-test discharges were calculated by the following equation:

$$q = 20.5\ R/t \tag{4}$$

where

q = the discharge rate, in barrels/day (to obtain the discharge rate in gallons per minute, a constant of 0.60 was used instead of 20.5);

R = length of fluid-filled drill stem, in feet; and

t = the flow period, in minutes.

The constant in equation 4, 20.5, assumes a standard drill-stem diameter of 3.83 in (Bredehoeft, 1965) and is equal to 0.01422 barrels per foot × 1,440 minutes per day; the constant, 0.60, is based on the assumption of 42 gallons per barrel. In order to use the drill-stem test to calculate permeability, it was decided arbitrarily in this investigation that the recovered fluid had to be at least 75 percent water or some form of water, such as muddy, salty, oil-cut, or gas-cut water.

Where data were insufficient for a Horner plot, summary information for the test was used to estimate local-scale permeability. One hundred forty-seven values of local-scale permeability were obtained from summary data by using two different methods.

If the information from two flow periods and two shut-in periods was available, local-scale permeability was determined by the following equation:

$$k = \frac{162.6\ qv\ \log[(\Delta t_1/\Delta t_2) \times (t_2 + \Delta t_2)/(t_1 + \Delta t_1)]}{h\left(p_{w_1} - p_{w_2}\right)} \tag{5}$$

where

t_1 = the initial flow period, in minutes;

t_2 = the final flow period, in minutes;

Δt_1 = the initial shut-in period, in minutes;

Δt_2 = the final shut-in period, in minutes;

p_{w_1} = the reported pressure during the initial shut-in period, in pounds per square inch; and

p_{w_2} = the reported pressure during the final shut-in period, in pounds per square inch; and all other variables are the same as in equation 1.

Equation 5 was derived from equation 1 by substitution of variables for each flow period and shut-in period separately into the original equation, subtraction to eliminate p from the equation, rearrangement of variables, and solution of simultaneous equations. An evaluation of equation 5 using data from actual drill-stem tests indicated that the equation gives reasonable results only if the following criteria are met:

1. The initial flow period should be at least 3 to 5 minutes long;
2. The second flow period should not differ from the first by more than 150 minutes;
3. The ratio of the initial shut-in period to flow period must be significantly different from the ratio of the final shut-in period to flow period; and
4. The initial shut-in pressure must exceed the final shut-in pressure.

If the summary information for a drill-stem test failed to meet these criteria, the test was not used to calculate permeability by this "two-flow-period" method.

Drill-stem tests with one flow period or an initial flow period of less than 3 to 5 minutes and decreasing shut-in pressure could be analyzed from summary information using an equation given by Earlougher (1977):

$$k = \frac{162.6\ qv\ \beta\ \log[(t + \Delta t)/\Delta t]}{h\left(p_{w_1} + p_{w_2}\right)} \tag{6}$$

where

β = a formation constant ranging from 0.99 to 1.06 (β can be ignored if unknown because it approximately equals 1); and all other variables are the same as in previous equations.

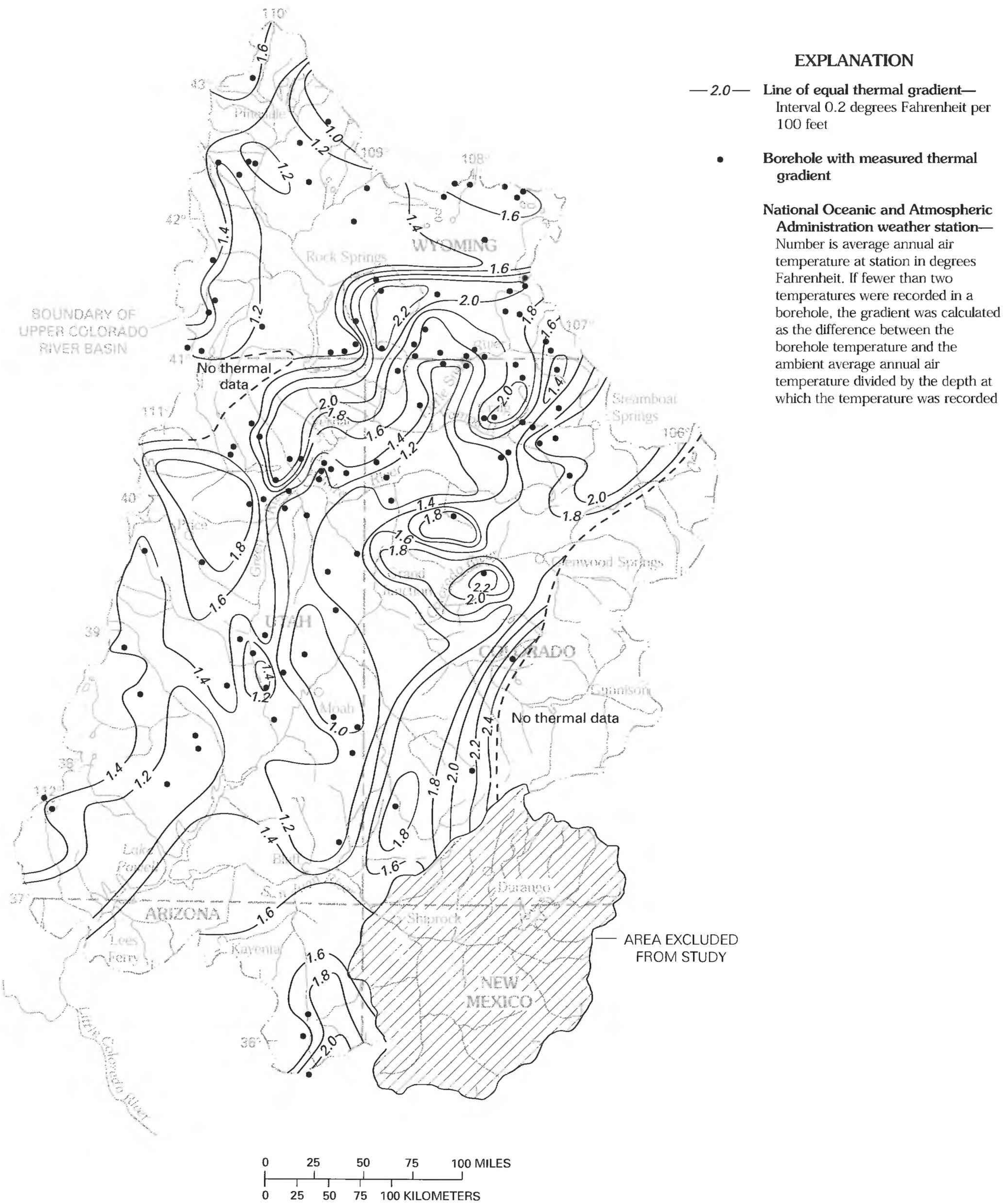


FIGURE 2.—Generalized distribution of thermal gradients in deep boreholes in the Upper Colorado River Basin.

In addition to local-scale permeability values obtained from drill-stem tests, 182 values of local-scale permeability were estimated from pore-scale permeability on the basis of equations developed during this study. These equations were determined from paired values of core and drill-stem test permeability values obtained from 22 stratigraphic intervals penetrated by either a single borehole or two boreholes no more than 5 mi apart (table 4). If the data from two boreholes were used, assurance that the data came from the same stratigraphic interval was based on the lithology of the formation and the apparent dip of stratigraphic markers between the boreholes. For each of the stratigraphic intervals, the drill-stem test permeability was plotted against the maximum value of core permeability and the arithmetic and geometric means of the core permeability values independent of rock type and by rock type. The largest correlation coefficients resulted when rock type was considered. For sandstone, the drill-stem test permeability was found to be related most closely to the geometric mean of the core permeability values (fig. 3*A*). For carbonate rocks, the drill-stem test permeability was found to be related most closely to the arithmetic mean of the core permeability values (fig. 3*B*).

On the basis of figure 3, local-scale permeability can be estimated from pore-scale permeability by the following equations:

$$k_1 = 1.4\, k_s^{0.65} \tag{7}$$

$$k_1 = 3.5\, k_c^{0.80} \tag{8}$$

where

k_1 = local-scale permeability, in millidarcies,
k_s = pore-scale permeability for sandstone, in millidarcies; and
k_c = pore-scale permeability for carbonate rocks, in millidarcies.

Because equations 7 and 8 are based on small numbers of paired values, their use is dependent on required precision. They were considered sufficiently reliable where used in the UCRB-RASA.

HYDRAULIC CONDUCTIVITY

Hydraulic-conductivity values used in this study were determined either by calculation from permeability values, by analysis of pressure-injection tests, or by calculation from transmissivity values. Transmissivity was determined either by constant rate, airlift, or step-drawdown pumping tests, bailing tests, slug tests, or flowing-well tests.

Five hundred eighty-eight values of hydraulic conductivity were calculated from permeability data. The permeability data consisted of 405 determinations from drill-stem tests, 135 estimates from laboratory-determined (pore-scale) permeability, and 47 estimates from porosity determined from borehole geophysical logs. One value of hydraulic conductivity was estimated from pressure-recovery data obtained during a flowing-well test of the Leadville Limestone near McCoy, Colo. Hydraulic conductivity was calculated from the flowing-well test after first estimating local-scale permeability using equation 5 (the two-flow period equation).

In this study, hydraulic conductivity was calculated from permeability by one of three equations, depending on the amount of information available for determining properties of the recovered fluid. If the temperature of the fluid was known, then viscosity could be estimated using equation 2 or 3. If the dissolved-solids concentration also was known, then hydraulic conductivity was calculated from the following equation (Weiss, 1982):

$$K = \frac{k \times \left[1 + \dfrac{TDS/1{,}000}{300}\right]}{365\nu} \tag{9}$$

where

K = hydraulic conductivity, in feet per day;
k = local-scale permeability, in millidarcies;
ν = viscosity, in centipoise; and
TDS = dissolved-solids concentration, in milligrams per liter.

If only the temperature of the fluid was known or could be estimated, the following equation was used:

$$K = k/365\nu \tag{10}$$

where all variables are the same as in equation 9. If neither the temperature nor the dissolved-solids concentration was known, the following equation (modified from Teller and Chafin, 1986, p. 9) was used:

$$K = k/410.5 \tag{11}$$

where

K = hydraulic conductivity at 60°F, in feet per day; and k is the same as in equation 9.

The factor, 410.5, is based on the fact that rock with a permeability of 1 millidarcy can transmit water at a rate of 0.00243 $ft^3/d/ft^2$ at 60°F.

Hydraulic conductivity was calculated directly from the results of 210 pressure-injection tests in 16 wells at 6 damsites. The data from these tests, and, in most cases, the calculated hydraulic-conductivity values were made available by the

TABLE 4.—*Comparison of closely obtained drill-stem test and core-permeability values for Paleozoic rocks in the Upper Colorado River Basin*

[Compared sites are no more than 5 miles apart; equivalent intervals identified by lithology and apparent dip of formation tops; dashes indicate not applicable]

Geologic unit	Drill-stem test site	Interval (in feet below land surface)		Drill-stem test permeability (in millidarcies)	Core site	Interval (in feet below land surface)		Arithmetic mean of values within interval	Geometric mean of values within interval	Maximum value within interval	Lithology of interval
		Top	Bottom			Top	Bottom				
Cutler Formation	NMB38–15–21bcb	3,628	3,655	1.4	NMB38–15–21bcb	3,629	3,651	7.9	3.2	40	Arkosic sandstone
Tensleep Formation	SB26–89–06ca	6,418	6,472	.13	SB26–90–01cda	6,338	6,391	4.7	.52	51	Quartz sandstone
Weber Sandstone	SB04–103–32cc	1,140	1,180	1.7	SB03–104–12bba	9,303	9,343	2.3	.41	22	Quartz sandstone
Weber Sandstone	SLA03–25–28bb	11,675	11,885	.14	SLA03–24–22bda	9,044	9,264	.41	.022	9.0	Quartz sandstone
Weber Sandstone	SLD05–22–23cba	4,047	4,169	17	SLD05–22–23cba	4,053	4,166	68	31	380	Quartz sandstone
Hermosa Formation	GA41–28–11ca	4,876	5,040	13	GA41–28–03cab	4,775	4,787	23	--	115	Dolomite
Hermosa Formation	GA41–30–16cdb	5,000	5,051	1.6	GA41–30–21cab	5,074	5,117	.18	--	1.0	Limestone
Hermosa Formation	NMB33.5–20–16cba	5,843	5,863	.44	NMB33.5–20–16cba	5,843	5,863	.62	--	4.9	Limestone and dolomite
Hermosa Formation	NMB33.5–20–25cb	5,800	5,850	1.1	NMB33.5–20–22bc	5,694	5,732	2.1	--	45	Limestone and dolomite
Hermosa Formation	NMB33.5–20–25cb	5,850	5,902	4.0	NMB33.5–20–21dac	5,721	5,771	9.8	--	60	Mostly dolomite
Hermosa Formation	SLD38–25–05dda	5,392	5,452	1.2	SLD37–25–32cad	5,665	5,675	.17	--	.38	Anhydrite and shaly dolomite
Hermosa Formation	SLD40–21–12dd	5,623	5,751	.16	SLD40–21–14aaa	5,370	5,414	.11	--	1.7	Limestone and dolomite
Hermosa Formation	SLD40–21–14aaa	5,170	5,251	17	SLD40–21–14aaa	5,231	5,250	13	--	120	Limestone and dolomite
Hermosa Formation	SLD40–23–14caa	5,627	5,700	30	SLD40–23–13bdd	5,696	5,751	3.1	--	47	Limestone
Hermosa Formation	SLD40–24–07bdc	6,165	6,240	.70	SLD40–24–22aba	5,893	5,919	.06	--	.43	Limestone
Hermosa Formation	SLD40–25–23cb	5,931	5,953	28	SLD40–25–26bbb	5,953	5,979	9.3	--	60	Limestone
Hermosa Formation	SLD41–23–12dd	5,557	5,606	200	SLD41–23–12aaa	5,592	5,642	12	--	146	Mostly limestone
Hermosa Formation	SLD41–25–26cc	5,364	5,394	14	SLD41–25–32bbd	5,347	5,377	1.2	--	27	Limestone and anhydritic limestone
Redwall Limestone	SLD23–17–16da	8,682	8,814	32	SLD23–17–17ada	8,693	8,824	6.3	--	73	Dolomite
Redwall Limestone	GA41–28–03bd	5,569	5,693	13	GA41–28–04dab	5,702	5,720	3.0	--	13	Cherty dolomite
Elbert Formation	GA37–29–35bbc	3,415	3,483	6.0	GA37–29–35bbc	3,442	3,466	8.9	1.1	24	Dolomitic and quartzitic sandstone
Elbert Formation	GA41–28–03bd	5,945	6,032	1.5	GA41–28–04dab	5,974	5,988	23	1.0	179	Sandstone

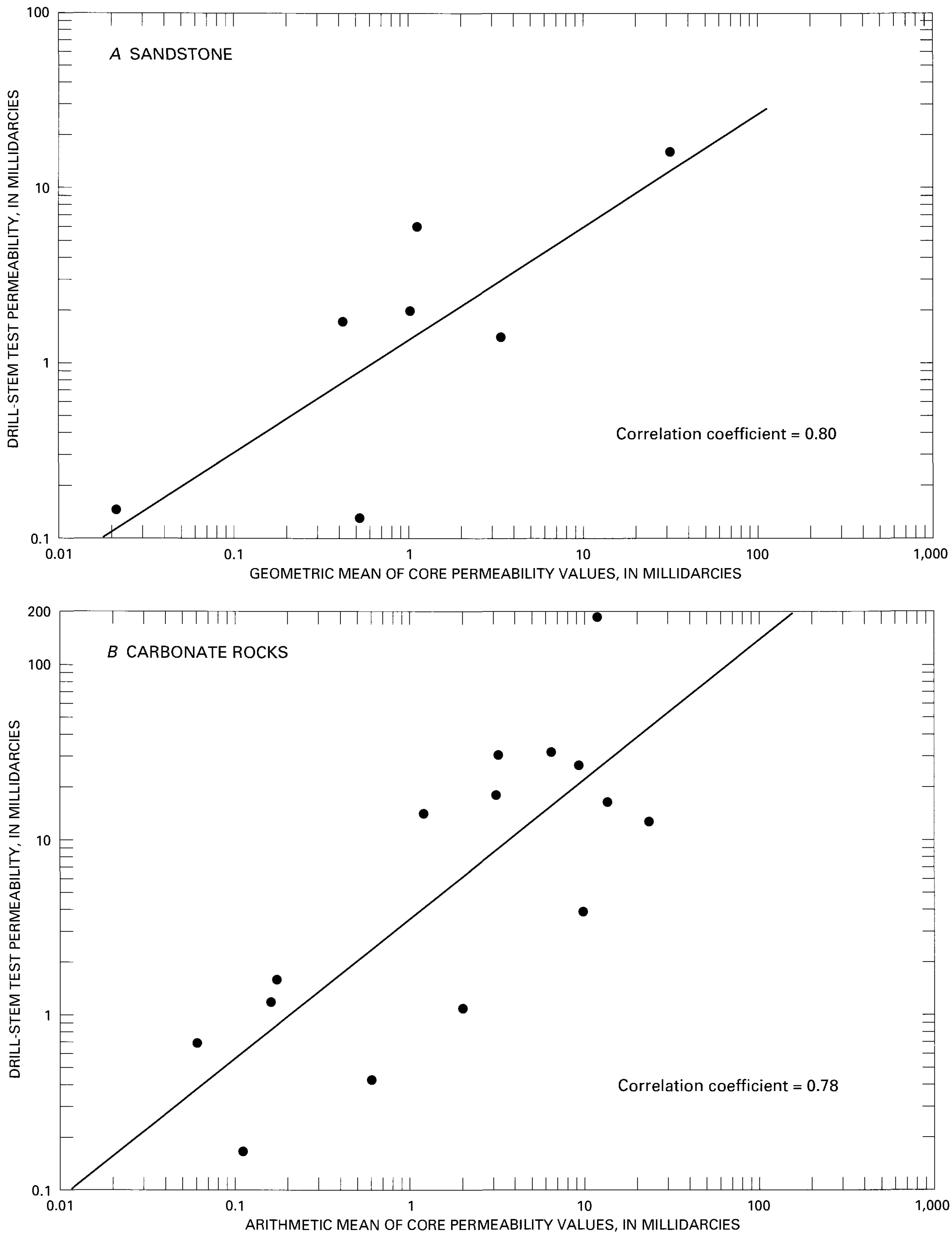


FIGURE 3.—Relations for sandstone and carbonate rocks between permeability determined from cores and permeability determined from drill-stem tests at closely spaced sites in the Upper Colorado River Basin.

Bureau of Reclamation (Salt Lake City and Denver offices, written commun., 1983–1985). The methods and analytical procedures used in the tests are described by Bureau of Reclamation (1981b, p. 249–266). The relevant equation is:

$$K = \frac{30.6\ Q \ln(L/r)}{LH} \tag{12}$$

where

K = hydraulic conductivity, in feet per day;
Q = discharge, in gallons per minute;
L = test interval, in feet;
r = well radius, in feet; and
H = head, in feet.

Hydraulic conductivity was calculated from 63 pumping, bailing, airlift, injection, and slug tests by dividing the transmissivity value obtained from the test by the test-interval thickness (methods of determining transmissivity from these tests are described in the next section). In a flowing-well test of the Leadville Limestone at Glenwood Springs, Colo., the hydraulic conductivity was calculated by dividing the thickness of the Redwall-Leadville zone, rather than the thickness open to the well (28 percent of the zone), because water is believed to be transmitted to the well used in the test by faults and fractures that permeate the entire Redwall-Leadville zone.

Regional distributions of unit-averaged hydraulic conductivity in hydrogeologic units were determined by two methods. For hydrogeologic units with sufficient data, values of hydraulic conductivity for intervals representative of the lithology of the entire hydrogeologic unit at the sites where these values were obtained were contoured. For hydrogeologic units with sparse data, unit-averaged hydraulic conductivity was determined at grid centers on a 10-mi × 10-mi grid from the thickness of each rock type in the hydrogeologic unit at the grid center and the median values of hydraulic conductivity for the individual rock types in the hydrogeologic unit. These estimated values were then contoured.

TRANSMISSIVITY

Very few determinations of transmissivity by aquifer tests were found during an extensive search of published reports and files of governmental agencies. For this reason, aquifer tests of formations in the UCRB from areas beyond the boundaries of the UCRB were used to support hydrogeologic interpretations. Accordingly, 14 flowing-well tests of the Flathead Sandstone, Madison Limestone, and Tensleep Sandstone in the Bighorn Basin, about 115 mi northeast of the crest of the Wind River Mountains (Cooley, 1985), and 2 pumping tests of the Coconino Sandstone in the Lower Colorado River Basin, 80 to 100 mi south of the UCRB (Akers, 1964; Cooley and others, 1969), were incorporated into the database.

Within the UCRB, only 15 constant-rate pumping tests were found. These included one test in the Cutler Group and upper member of the Hermosa Formation (Thackston and others, 1984, p. 21–31), two tests in the Weber Sandstone (Hood, 1976, p. 53–54; Sumsion, 1976, p. 51–59), six tests in the De Chelly Sandstone (Cooley and others, 1969), and six tests in disturbed blocks (caprock) of the Paradox Member of the Hermosa Formation (Wollitz and others, 1982). The latter tests were given little weight in interpretation of the hydrologic properties of the Paradox Member because the data curves were not convincingly analyzable by the method used, and the transmissivity of the caprock probably differs from that of the rock in place. The 15 pumping tests were analyzed by either the Theis (1935) nonequilibrium solution, the Theis (1935) recovery method, or the Cooper and Jacob (1946) modification of the Theis nonequilibrium solution. Test procedures are described by Lohman (1979). The equations for the Theis nonequilibrium solution are:

$$T = \frac{Q}{4\pi s} W(\mu) \tag{13}$$

where

T = transmissivity, in feet squared per day;
Q = discharge, in cubic feet per day;
s = drawdown, in feet, at the type-curve match point; and
$W(\mu)$ = the nonleaky well function of μ at the type-curve match point.

$$\mu = \frac{r^2 S}{4Tt} \tag{14}$$

where

r = distance from the center of the well, in feet;
S = storativity, dimensionless;
t = time since flow started, in days; and
T is the same as in equation 13.

The equation for the Theis recovery and Cooper-Jacob methods, assuming $\mu \le 0.01$, is:

$$T = \frac{35.2Q}{\Delta s_d} \tag{15}$$

where

T = transmissivity, in feet squared per day;
Q = discharge, in gallons per minute; and
s_d = change in drawdown, residual drawdown, or recovery, in feet, over 1 log cycle of time.

An "airlift" pumping test of the Cutler Group and upper member of the Hermosa Formation (Woodward-Clyde Consultants, 1982; Thackston and others, 1984) provided one of only two composite transmissivity values for the entire thickness of the Cutler-Maroon zone of the Canyonlands aquifer. The test was done by displacing water from a well using compressed air and monitoring the recovery of head (pressure). Although not specifically stated, the recovery data presumably were analyzed using equation 15.

The data from an aborted step-drawdown pumping test of the Leadville Limestone at Ouray, Colo. (fig. 4), that was done in 1988 were analyzed by separating the total drawdown into components related to individual increments of discharge added during the test. For example, from 1 to 5 minutes, with the pump discharging at a rate of 49.6 gal/min, total drawdown remained constant at 4.44 ft. Therefore, it was assumed that the drawdown related to the first increment of discharge remained constant at 4.44 ft throughout the test. At 6 minutes, the discharge was increased by 49.7 gal/min, and the drawdown increased to 10.22 ft. Therefore, the drawdown attributed to the second increment of discharge (49.7 gal/min) at 6 minutes was 10.22 ft – 4.44 ft = 5.78 ft. Values of drawdown attributed to the second discharge increment from 6 to 12 minutes were plotted against the log of time, and a trend was extrapolated for further calculations. At 14 minutes, the discharge was increased by 51.7 gal/min, and the total drawdown was 16.00 ft. The drawdown attributable to the second increment of discharge at 14 minutes was determined from the extrapolated trend of drawdown with time to be 11.1 ft. Thus, the drawdown attributable to the third increment of discharge at 14 minutes was 16.00 ft – 4.44 ft –11.1 ft = 0.46 ft. Continuation of this process for all subsequent measurements of drawdown during the test allowed calculation of the drawdown with time attributable to each of the five increments of discharge added during the pumping test. The semilog plots of drawdown against time for all but the first discharge increment (fig. 5) were analyzed using equation 15, and the separate transmissivity values determined were averaged to obtain a solution for the test (discussed in the section on the Redwall-Leadville zone).

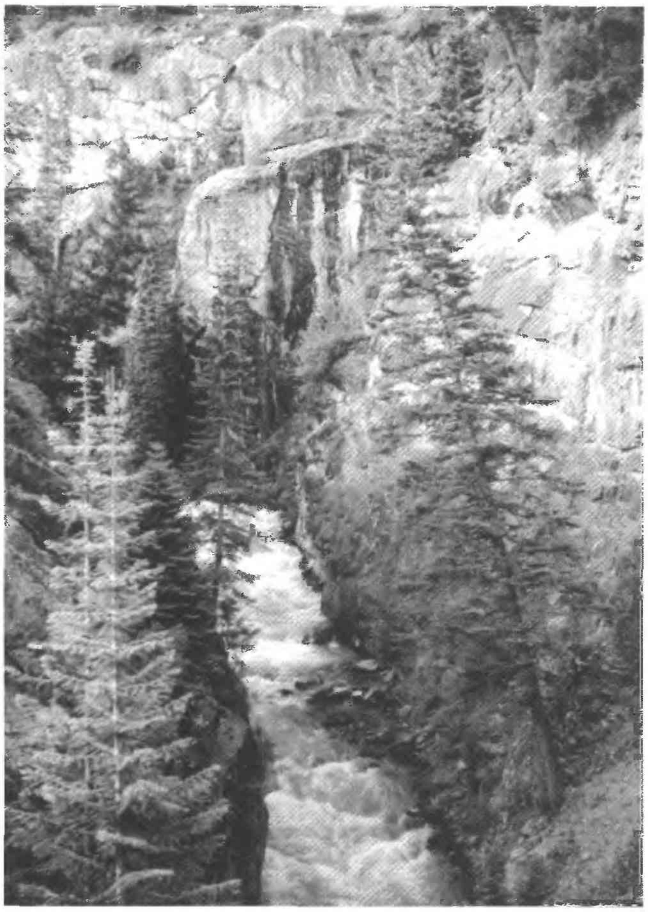


FIGURE 4.—Uncompahgre River at Ouray, Colorado, near site of well OX–3 (NMB44–07–31cbd$_3$). Cliffs above the river are Leadville Limestone of Mississippian age.

Two bailing tests in the Belden and Maroon Formations were analyzed by applying equation 15 to residual drawdown data from these tests; another bailing test in the Belden Formation was analyzed using the method of Skibitzke (1958). The data for these tests, all of which were done in 1963 during site investigation for Ruedi Dam near Meredith, Colo. (location shown on pl. 1), were provided by the Bureau of Reclamation (written commun., 1985). The Skibitzke (1958) equation is:

$$T = \frac{V}{4\pi s' t} \tag{16}$$

where

T = transmissivity, in feet squared per day;
V = volume bailed, in cubic feet;
s' = residual drawdown, in feet; and
t = time since bailing stopped, in days.

An "airlift" test of the Leadville Limestone at Ouray was done on July 15, 1987. All of the water in a 320-ft-deep, flowing well was blown out using compressed air during a 5-minute period. Recovery of the water level in the well then was monitored. Since the water in the well was removed essentially instantaneously, the test could be analyzed as a slug test using the method of Cooper and others (1967) and type curves prepared by Reed (1980).

Three slug tests done in salt beds of the Paradox Member of the Hermosa Formation by Rush and others (1980) provided additional transmissivity data. These tests also were analyzed by the method of Cooper and others (1967).

Thackston and others (1984) described four injection tests of the Redwall Limestone at a site on the Monument upwarp. Transmissivity and hydraulic conductivity were determined

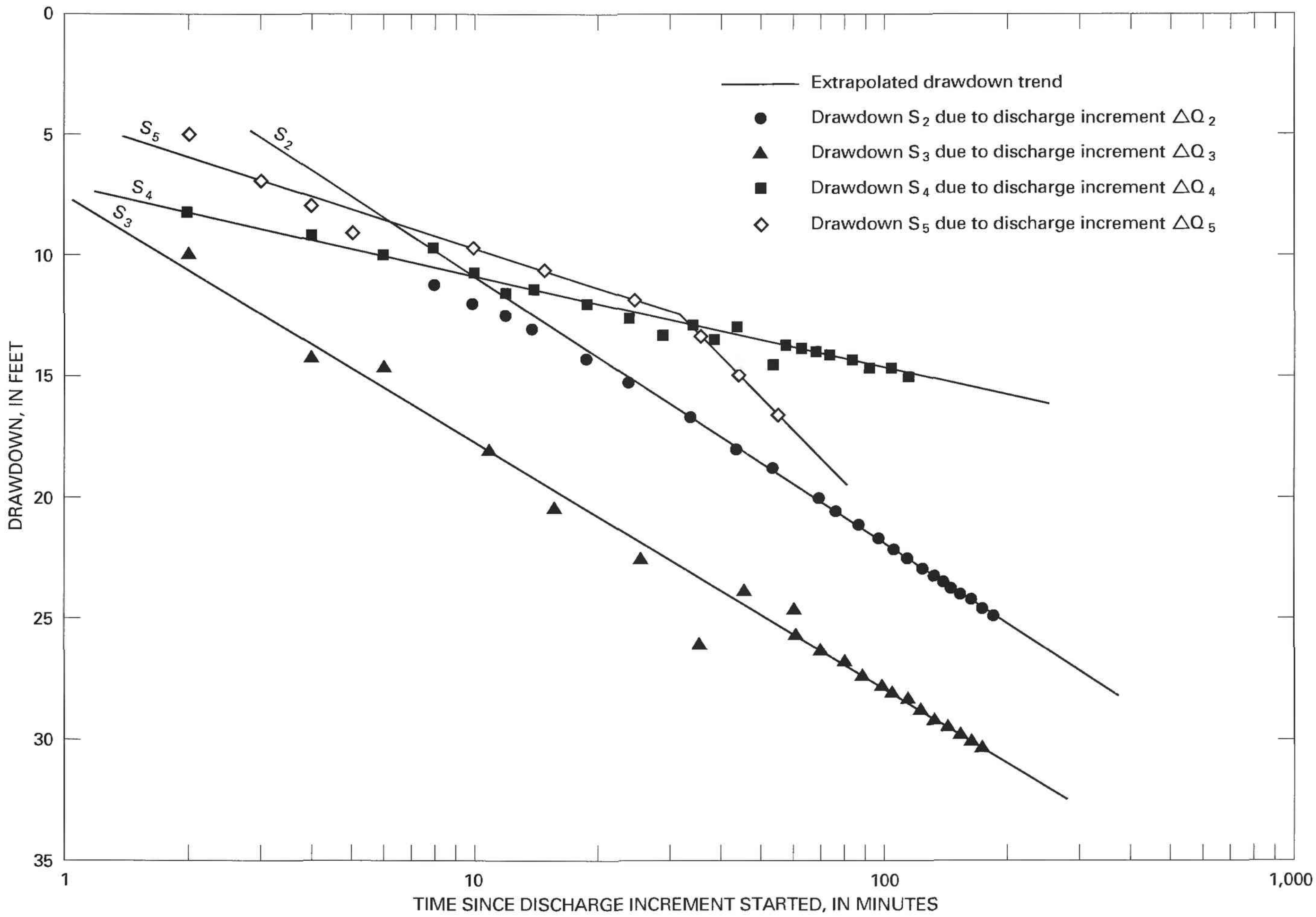


FIGURE 5.—Drawdown in well OX–3 (NMB44–07–31cbd$_3$) at Ouray, Colorado, during aquifer test of April 25, 1988. (Drawdown during the first increment of discharge did not change with time and is not shown. The drawdown during the fifth increment of discharge apparently was affected by a barrier boundary about 32 minutes after this increment began. Data from the test were provided by David Vince, City of Ouray, written commun., 1988.)

from these tests using equations 1 and 15. For each property, the calculated values from the four tests were so similar that the average was considered to be a reasonably good indication of the property being determined. These average values were added to the database.

Two flowing-well tests of the Leadville Limestone were analyzed. One of these, at McCoy, Colo., was analyzed using equation 5 to obtain local-scale permeability. Permeability was converted to hydraulic conductivity by equation 11 and to transmissivity by multiplying the hydraulic conductivity by the test-interval thickness.

A flowing-well test of the Leadville Limestone at Glenwood Springs, Colo. (Geldon, 1989c), was made during this study and analyzed using an extensive monitoring network. Changes in head in the production well at Glenwood Springs were observed directly by means of a standpipe attached to the well. Discharge was measured by the orifice-plate method (Bureau of Reclamation, 1981b, p. 233–242). Four observation wells and three springs were monitored at Glenwood Springs in addition to the production well. A recording barometer was set up near the site to adjust recorded water levels for changes in atmospheric pressure. Test data from the Glenwood Springs test were analyzed by the straight-line method of Jacob and Lohman (1952) and the type-curve matching method of Hantush (1960). The equation for the Jacob-Lohman method is:

$$T = \frac{0.183 \times 1{,}440}{\Delta(s/Q)_d} \qquad (17)$$

where

T = transmissivity, in feet squared per day; and

$\Delta(s/Q)_d$ = change in specific discharge over 1 log cycle of time, in feet per gallon per minute.

The equation for the Hantush (1960) type-curve matching method is:

$$T = \frac{192.5\, Q\, H(\mu)}{4\pi s} \qquad (18)$$

where
T = transmissivity, in feet squared per day;
Q = discharge, in gallons per minute;
s = drawdown or recovery at type-curve match point; in feet; and
$H(\mu)$ = the leaky aquifer with storage in the confining bed well function of μ at the type-curve match point; and
μ = the same as in equation 14.

Thirty-eight specific-capacity (pumping, bailing, and airlift) tests on file with the Colorado Department of Natural Resources, Division of Water Resources, Office of the State Engineer, provided the remaining transmissivity values. Transmissivity was estimated from specific capacity using either of two equations. If the well radius and pumping duration were known, then the following equation from Lohman (1979, p. 52) was used:

$$\frac{Q}{s_w} \approx \frac{4\pi T}{2.3\ \mathrm{Log}\left(\frac{2.25\ Tt}{r_w^{\ 2} S}\right)} \qquad (19)$$

where
Q = discharge, in cubic feet per day;
s_w = drawdown, in feet;
T = transmissivity, in feet squared per day;
t = pumping time, in days;
r_w = radius of well, in feet; and
S = storativity, dimensionless, and assumed equal to the test interval thickness multiplied by 0.000001.

If only the discharge and drawdown were known, then transmissivity was estimated using the following equation from Driscoll (1986, p. 1021):

$$T = \frac{2{,}000\ Q}{s_w} \qquad (20)$$

where all variables are the same as in equation 18, except that T is in gallons per day per foot.

Regional distributions of composite transmissivity in the hydrogeologic units were determined for each unit by contouring grid-center values on a 10-mi × 10-mi grid. If a measured or estimated value of transmissivity representative of the entire thickness of a hydrogeologic unit was near a grid center, it was used for the grid-center value. Otherwise, the grid-center value was obtained by multiplying the unit-averaged hydraulic conductivity by the thickness of the hydrogeologic unit at the grid center. Contouring of the grid-center values was guided by the structural setting.

STORATIVITY

Conventionally, storativity is calculated from observation-well drawdown or recovery data in a pumping or flowing-well test (see Lohman, 1979). Only the flowing-well test of the Leadville Limestone at Glenwood Springs done during this study provided data sufficient for the calculation of storativity. The equation used was:

$$S = \frac{4T\, t\, \mu}{1{,}440 r^2} \qquad (21)$$

where
S = storativity, dimensionless;
T = transmissivity, in feet squared per day;
t = time at type-curve match point, in minutes;
r = distance of observation well from production well, in feet; and
μ = a type-curve match point.

Specific storage equals the storativity divided by thickness. The regional range in storativity for the Redwall-Leadville zone of the Madison aquifer was estimated from specific storage in the Glenwood Springs area and the range in thickness of the Redwall-Leadville zone in the UCRB. Specific storage in the Glenwood Springs area was calculated from the storativity of the Leadville Limestone determined during the flowing-well test at Glenwood Springs and the thickness of the Redwall-Leadville zone at Glenwood Springs.

Storativity also can be estimated from porosity and known or assumed physical constants of water and rock, as described by Lohman (1979, p. 9):

$$S = \Phi\gamma b\left(\beta + \frac{\alpha}{\Phi}\right) \qquad (22)$$

where
S = storativity, dimensionless;
Φ = porosity, in percent;
γ = specific weight of water, which equals 0.036 pounds per cubic inch;
b = aquifer thickness, in feet;
β = compressibility of water, which equals 0.0000033 square inch per pound; and
α = compressibility of the rock skeleton, in square inches per pound.

The regional distribution of storativity in the Weber-De Chelly zone of the Canyonlands aquifer was estimated using equation 22, values of porosity and thickness determined on a 10-mi × 10-mi grid, and the compressibility of the Cretaceous Fox Hills Sandstone (0.0000011 in^2/lb), as reported by Taylor (1968).

YIELD

Yield values reported in this study, unless otherwise noted, are artesian flows into or from wells or from springs. The well flows include discharges measured at the surface and discharges from specific horizons recorded during drilling or aquifer tests. Much of the yield data was calculated from the volume of fluid recovered during drill-stem tests (using equation 4). If more than one flowing interval was penetrated in a borehole or in closely spaced boreholes, the largest yield was used in characterizing the hydrogeologic unit in the area where the data were obtained. More than 800 yield values were available for interpretation.

Maps showing regionally distributed yields indicate the yield to be expected from the most productive horizon in a hydrogeologic unit under artesian (natural flow) conditions. Larger yields might occur if more than one producing horizon were penetrated by a well. Larger yields also could be obtained by pumping at a rate exceeding the natural flow rate (which would cause the water level in the well to be lowered). Smaller yields might occur if a well did not penetrate the most productive horizon or penetrated only a fraction of the thickness of this horizon.

For each hydrogeologic unit, determination of the probability of obtaining a certain yield was biased not only by the method of selecting data for this study but also by selectiveness in the original yield determinations. First, only wells drilled for a water supply that produced usable quantities of water were included in the database. Second, only drill-stem tests that produced enough water to calculate a discharge were included in the database. Third, drill-stem tests preferentially were conducted in intervals with a potential for petroleum production; less permeable intervals routinely were not tested, and intervals capable of yielding large quantities of water but not economically producible quantities of petroleum might have been overlooked. Finally, where springs were observed, generally only flows from the largest springs were measured or reported. Taking into account the nonrigorous sampling procedures, the true probability of a particular yield from a hydrogeologic unit at any site may differ from what is indicated by maps in this report.

PHYSICAL SETTING

In the UCRB, recurrent tectonic activity from Precambrian to Quaternary time, a period of more than 600 million years, has molded the terrain into extremely diverse landforms. Precipitation varies with the topography. The combination of topography and precipitation determines the course and discharge of the region's streams, and all of the above factors affect where water is added to or discharged from the Paleozoic rocks. Circulation of water in the Paleozoic rocks, therefore, not only is a function of the lithologic and hydrologic properties of these rocks but also is a reflection of the past and present effects of structural disturbance and erosion and the vagaries of a climate that is ever changing in both time and space.

TOPOGRAPHY

In the UCRB, the forces of running water, snow and rain, glaciers, and wind have carved a tremendous thickness of sedimentary, igneous, and metamorphic rocks into a jumbled landscape of mountains, plateaus, plains, and valleys. At the confluence of the Green and Colorado Rivers in Utah, the geologist and explorer, John Wesley Powell (1895) wrote,

> Wherever we look there is but a wilderness of rocks—deep gorges where the rivers are lost below cliffs and towers and pinnacles, ten thousand strangely carved forms in every direction, and beyond them mountains blending with the clouds.

The UCRB includes parts of four physiographic provinces—the Middle Rocky Mountains, Wyoming Basin, Southern Rocky Mountains, and Colorado Plateaus (pl. 1). The land surface generally lies between altitudes of 5,000 to 8,000 ft (above the NGVD of 1929), but mountains rise as high as 12,000 to 14,500 ft in altitude, and canyons 1,000 to 3,000 ft deep are common. The lowest point in the area, where the Colorado River leaves the basin at Lees Ferry, Ariz., is at an altitude of about 3,100 ft.

MIDDLE ROCKY MOUNTAINS

The Middle Rocky Mountains physiographic province in the UCRB and vicinity includes the Wind River Mountains, Gros Ventre Range, Overthrust Belt, Uinta Mountains, and Yampa Plateau (pl. 1). The northwest-trending Wind River Mountains (fig. 6*A*) rise abruptly above the Green River Basin to the southwest to altitudes in excess of 13,500 ft. Dissected erosional surfaces that cap the Wind River Mountains descend gradually northeastward to a series of cuestas and hogbacks flanking the Wind River Basin. The Gros Ventre Range is a northwest-trending but subdued extension of the Wind River Mountains. The Overthrust Belt consists of several north- to northwest-trending, steep-sided mountain ranges, ridges, and intermontane valleys. Mountain crests exceed 11,000 ft in the northern part of the area but decrease in altitude southward; in the southern part of the area, mountains give way to ridges, which rise only 500 to 1,000 ft above adjacent lowlands. The Uinta Mountains and Yampa Plateau, together, form an unusual east-trending range. In the western Uinta Mountains, a sinuous arete with many spurs and pyramidal peaks 12,000 to 13,500 ft in altitude rises 1,000 to 2,000 ft above a broad-crested upland surface (fig. 6*B*). In the eastern Uinta Mountains, Browns Park and narrow canyons of the Yampa and Green Rivers segment the range into four plateaus—Cold Spring Mountain, Diamond Mountain, Douglas Mountain, and Blue Mountain. Plateau surfaces range from 8,000 to 9,700 ft in altitude.

WYOMING BASIN

The Wyoming Basin physiographic province in the UCRB and vicinity includes the Hoback, Green River, Great Divide, and Washakie Basins (collectively referred to as the greater

A

B

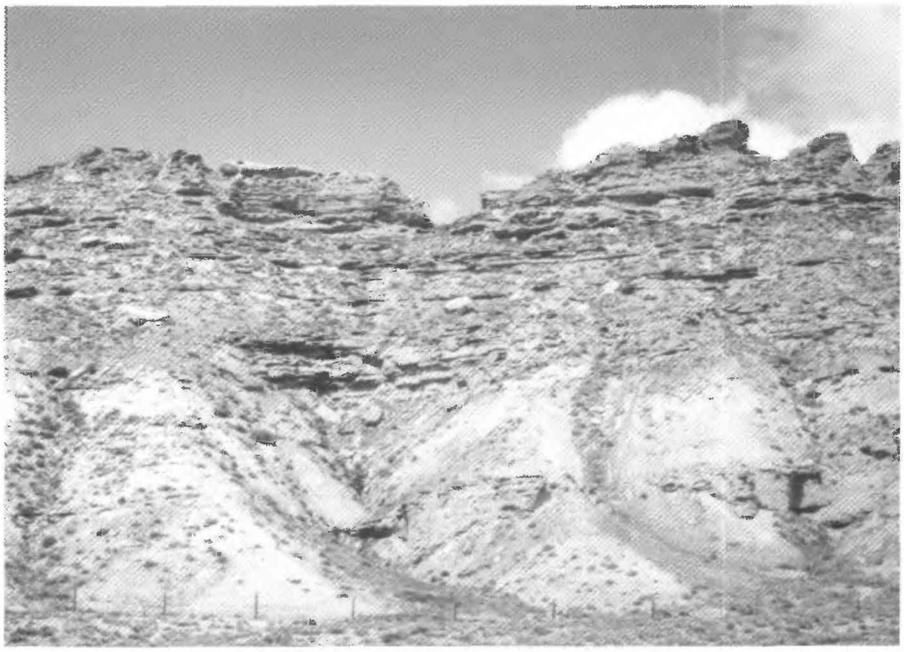
FIGURE 6.—Landscapes of the Middle Rocky Mountains physiographic province: *A*. Wind River Mountains—Square Top Mountain, composed of Archean gneiss, is reflected in Upper Green River Lake near the headwaters of the Green River north of Pinedale, Wyoming. *B*. Western Uinta Mountains—Mount Agassiz, composed mostly of quartzite of the Proterozoic Uinta Mountain Group, towers above a meadow northeast of Kamas, Utah.

Green River Basin); the Sand Wash and Hanna Basins; the Sweetwater Arch; and the Rock Springs and Rawlins Uplifts (pl. 1). All of the basins are characterized by undulating, sage-covered plains and badlands locally rising to low hills, buttes, mesas, and isolated mountains (fig. 7*A*). The floors of the basins generally lie at altitudes between 6,000 and 7,500 ft. However, cuestas bordering the Rock Springs and Rawlins Uplifts (fig. 7*B*) and scattered buttes, such as the Oregon Buttes, crest at altitudes between 7,500 and 8,500 ft, and in the Sand Wash Basin, the Williams Fork and Elkhead Mountains tower 2,000 to 5,000 ft above adjacent lowlands.

SOUTHERN ROCKY MOUNTAINS

The Southern Rocky Mountains physiographic province in the UCRB and vicinity includes the Sierra Madre; Park, Gore, Rabbit Ears, Front, and Sawatch Ranges; Elk, West Elk, and

A



B

FIGURE 7.—Landscapes of the Wyoming Basin physiographic province: *A*. Green River Basin—Bluffs underlain by the Tertiary Wasatch and Green River Formations near La Barge, Wyoming. *B*. Rock Springs Uplift—A cuesta of Cretaceous sandstone and shale flanks the eroded interior of the uplift known as the Baxter Basin, east of Rock Springs, Wyoming.

San Juan Mountains; White River and Gunnison Plateaus; and the Middle Park and Eagle Basins (pl. 1). In the words of Frank Waters (quoted in Boddie and Boddie, 1984, p. 44),

> With all their infinite variations, the mountains comprise not only heaving waves of forest, but jutting cliffs, abysmal gorges, and deep sunless canyons, vast open parks and tiny arctic meadows, small blue lakes, gushing warm geysers, mineral springs, cold trout pools, lacy falls, heavy cataracts and great soggy marshes, cones and craters of extinct volcanoes, bristling hogbacks, rolling hills of sage and cedar, high groves of aspen, immense flat-topped mesas, solitary bluffs, and weirdly eroded buttes.

The Southern Rockies are the backbone of the continent, giving rise to four great rivers—the Platte, Rio Grande, and Arkansas, which flow to the Atlantic, and the Colorado, which

A

B

FIGURE 8.—Landscapes of the Southern Rocky Mountains physiographic province: *A*. Elk Mountains—Two 14-thousand-foot peaks, Maroon Peak (14,156 feet) on the left and Pyramid Peak (14,018 feet) on the right, flank an alpine basin at the head of Maroon Creek, between Aspen and Crested Butte, Colorado. All of the rocks exposed belong to the Pennsylvanian and Permian Maroon Formation. The view is north from West Maroon Pass. *B*. Sawatch Range—Mount Harvard (14,420 feet) and neighboring peaks composed of Proterozoic igneous and metamorphic rocks. View is southeast from Mount Belford (14,197 feet).

flows to the Pacific. Fifty-four peaks (twenty-nine in the UCRB) in the San Juan Mountains, Elk Mountains (fig. 8*A*), Gore Range, Sawatch Range (fig. 8*B*), and Front Range exceed 14,000 ft in altitude. Peaks of 12,000 to 13,000 ft in altitude are so common that many are unnamed.

Below the mountain summits, the land descends in a series of dissected erosional surfaces to high plateaus and intermontane basins. The White River Plateau is a broad, lava-capped highland that is bordered on the west by the sinuous Grand Hogback Monocline and transected near its southern end by Glenwood Canyon, a 1,800-ft-deep gorge carved by the Colorado River. The Gunnison Plateau is a narrow upland underlain by Precambrian crystalline rocks into which the Gunnison River has incised a 2,000-ft-deep gorge known as the Black Canyon. Surfaces of the White River and Gunnison Plateaus generally lie at altitudes between 9,000 and 12,000 ft. In the Eagle Basin and Middle Park, streams meander through wide alluvial valleys bordered by terraces, low hills, and isolated mountains. Surface altitudes in these basins generally range from 5,000 to 8,000 ft.

COLORADO PLATEAUS

The Colorado Plateaus province is the largest and most topographically diverse region in the UCRB. The outstanding topographic features include high, forested plateaus; broad, inwardly sloping, dissected plateaus with intricately eroded escarpments; terraced plateaus and plains studded with buttes, mesas, arches, and spires; eroded domes ringed by cuestas; flat-bottomed valleys flanked by parapet-like ridges; snowcapped, laccolithic mountains; and, everywhere, tortuous, vertically walled canyons (fig. 9). As Powell (1895) wrote,

> Every river entering *** has cut a canyon; every lateral creek has cut a canyon; every brook runs in a canyon; every rill born of a shower and born again of a shower and living only during these showers has cut for itself a canyon; so that the whole [area] is traversed by a labyrinth of these deep gorges.

The Colorado Plateaus physiographic province in the UCRB and vicinity includes the Uinta, Piceance, Henry Mountains, Kaiparowits, Kaibito, Blanding, and Black Mesa Basins; the Uncompahgre, Defiance, Kaibab, Coconino, and San Francisco Plateaus; the High Plateaus region of Utah, which includes the Paunsaugunt, Table Cliffs, Aquarius, Awapa, Fish Lake, and Wasatch Plateaus; the Capitol Reef and Paradox Basin Fold and Fault Belts; the Castle Valley, San Rafael Desert, Grand Valley–Montrose Valley, and Gallup Sags; the Marble and Four Corners Platforms; and the San Rafael Swell, Circle Cliffs Uplift, Monument Upwarp, Tyende Saddle, Douglas Creek Arch, Chuska Mountains, Echo Cliffs, Preston Bench, and Mogollon Slope (pl. 1). Six groups of laccolithic mountains—the Abajo, Henry, and Carrizo Mountains, Sleeping Ute Mountain, Navajo Mountain, and La Sal Mountains—are scattered throughout these areas. Although the Axial Basin Arch continues the structural trend between the Middle and Southern Rocky Mountains (Stone, 1986, fig. 1) and probably should be included in one of those physiographic provinces, traditionally, and in this report, it is considered part of the Colorado Plateaus province.

The Colorado Plateaus province, in general, lies between altitudes of 5,000 and 7,000 ft; but, as described by Gregory and Moore (1931, p. 12–14):

> *** the downward departures *** are approximately equal in amount to the upward departures ***. For the region as a whole, changes in altitude are abrupt; gentle slopes are conspicuously absent. Above the valley floors the plateau benches rise by steps, bench after bench, and into the benches the streams are sunk an equal amount ***. The plateau benches are so continuous that canyons which cut their edges appear at a distance as insignificant breaks in a horizontal skyline ***.

Surfaces of the Uncompahgre, Defiance, and Kaibab Plateaus; the High Plateaus region of Utah; and the Chuska Mountains generally lie between altitudes of 7,500 and 11,500 ft; escarpments of the inwardly sloping broad plateaus that comprise the Piceance, Uinta, Kaiparowits, and Black Mesa Basins crest at altitudes between 7,000 and 11,000 ft; peak altitudes of laccolithic mountains and isolated erosional and structurally raised mountains, such as Juniper Mountain, Cross Mountain, the Danforth Hills, Elk Ridge, Thousand Lake Mountain, and Canaan Peak, range from 8,000 to 13,000 ft. Canyons 1,000 to more than 3,000 ft deep include Unaweep Canyon and Ute Canyon in the Uncompahgre Plateau, Canyon De Chelly in the Defiance Plateau, and numerous canyons of the Colorado, Green, San Juan, Dolores, San Miguel, Dirty Devil, San Rafael, Escalante, and Paria Rivers.

CLIMATE

The climate of the UCRB varies from arid in the lowlands that are in the rain shadow of mountains and plateaus forming the western boundary to humid in the mountains and higher plateaus. The average annual precipitation ranges from less than 6 to more than 40 inches (fig. 10), generally increasing with increasing altitude (fig. 11). In the part of the UCRB that does not include the San Juan Basin, the average annual precipitation is about 15 inches (as discussed later in this report).

Average precipitation amounts can be misleading in arid and semiarid parts of the UCRB, where monthly accumulations recorded at weather stations can be the result of a single storm or a few downpours of local extent (Gregory, 1938, p. 15). In these areas, average annual precipitation can vary considerably from year to year. At Moab, Utah, for example, the average annual precipitation is 8.18 inches, but recorded precipitation has ranged from 3.02 inches in 1956 to 15.96 inches in 1918 (Sumsion, 1971, p. 7). At Scofield Dam, in the Book Cliffs region of Utah, the average annual precipitation is 16.0 inches, but precipitation ranged from 6.77 to 32.03 in/yr between 1931 and 1975 (Waddell

A

B

FIGURE 9.—Landscapes of the Colorado Plateaus physiographic province: *A*. Monument Upwarp—looking southwest from the Island in the Sky district. In the foreground is the White Rim Bench, which is capped by the White Rim Sandstone and underlain by the Organ Rock Shale, both of which are in the Permian Cutler Group (of Baars, 1962). In the distance, cliffs of Jurassic Wingate Sandstone, Kayenta Formation, and Navajo Sandstone rise above the Green River at the entrance to the Maze. *B*. Paradox Basin southeast of Moab, Utah. La Sal Mountains, underlain by Tertiary igneous rocks, rise above cliffs of Jurassic Entrada Sandstone.

and others, 1981, p. 5). The consequence of these observations is that one cannot necessarily eliminate the possibility of ground-water recharge occurring in areas with small average annual precipitation because in some years, the annual precipitation or the accumulation from individual storms in these areas may be so large that some precipitation may infiltrate to the water table.

Precipitation in the UCRB predominantly results from eastward moving cyclonic storms in the winter months and northeasterly moving convection cells in the summer months (Gregory, 1938, p. 15; Iorns and others, 1965, p. 10). The cyclonic storms (which originate in the Pacific Ocean) bring prolonged rain or snow, whereas the convection storms (which originate in the Gulf of Mexico) produce brief but intense localized rain. Reflecting these dichotomous sources of precipitation, many areas have dual precipitation peaks—one between December and March and a second between July and October. Examples include the Paradox Basin (Rush and others, 1982), Uinta Mountains (Hood and Fields, 1978), Uinta Basin (Price and Miller, 1975), San Juan Mountains (Eckel and others, 1949), Capitol Reef Fold and Fault Belt (Smith and others,

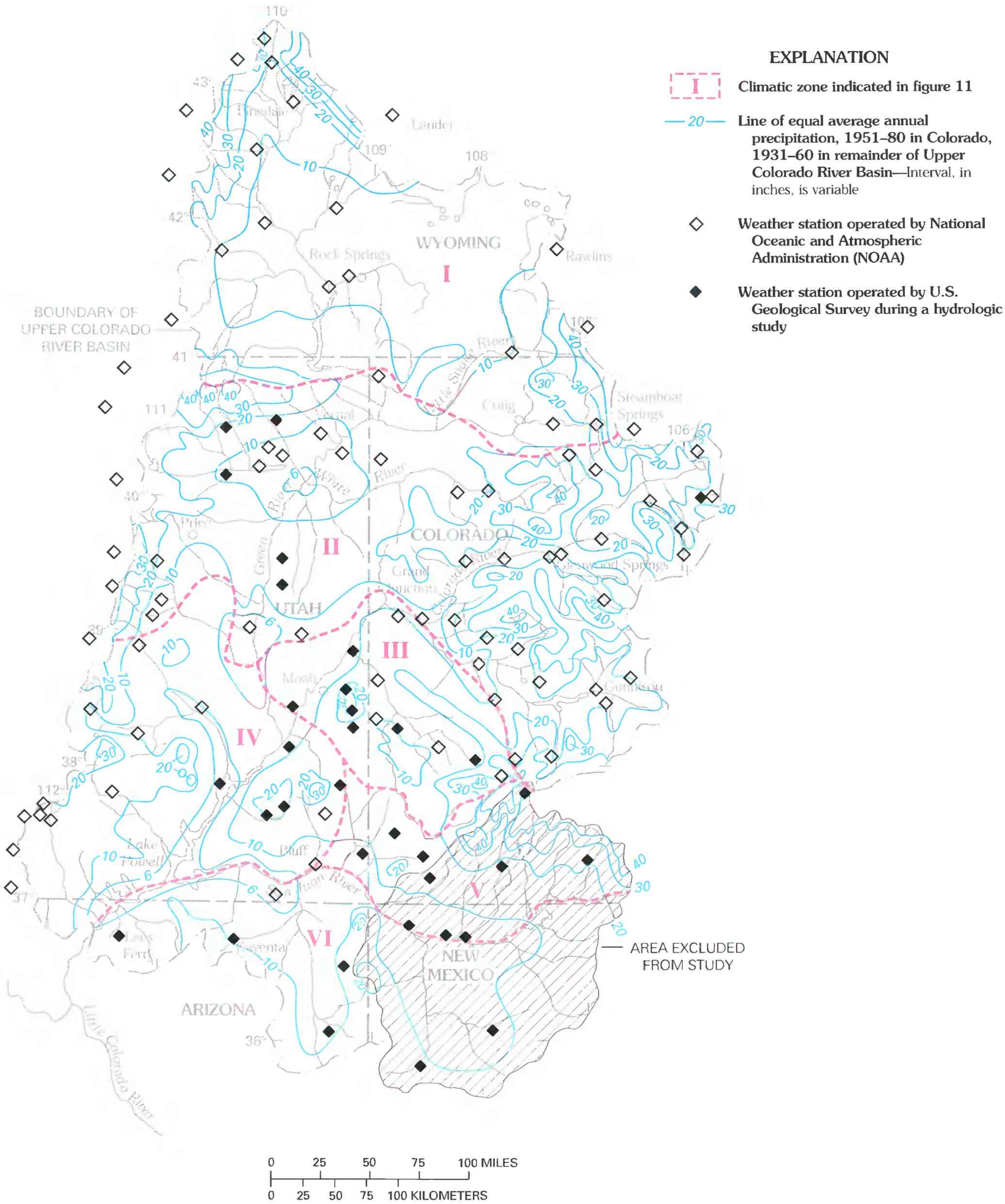


FIGURE 10.—Average annual precipitation in the Upper Colorado River Basin and location of sites used in analyses of climatic data. (Modified from Price and Arnow, 1974; Doesken and others, 1984.)

1963), Navajo-Hopi Reservations (Cooley and others, 1969), and the High Plateaus region of Utah (Gregory, 1951; Plantz, 1985). However, some areas, such as the Four Corners Platform (Irwin, 1966) and Monument Valley (Gregory, 1938), have a single precipitation peak, which occurs between July and October. The minimum monthly precipitation often occurs in early autumn or late spring but can occur at any time during the year from site to site within a geographic area (National Oceanic and Atmospheric Administration, 1930–85).[1]

A substantial amount of precipitation is consumed by evapotranspiration. Evapotranspiration rates generally decrease inversely with altitude as the temperature decreases (fig. 11) but, nevertheless, are large throughout much of the study area. For example, in the Cottonwood Wash watershed near Cortez, Colo., the average annual precipitation is 17 inches; evapotranspiration in this area was estimated, on the basis of climate statistics and physiographic features, to range from 60 to 90 percent of the annual precipitation from 1979 to 1982 (Geldon, 1985a, p. 10–14). At Black Mesa, in northeastern Arizona, Eychaner (1983, p. 9–11) indicated that as much as 46 percent of the ambient precipitation on the Jurassic Navajo Sandstone is consumed by evapotranspiration. At Scofield Dam, in the Book Cliffs region of Utah, measured lake evaporation exceeds the average annual precipitation by 26 in/yr (Waddell and others, 1981, p. 6). Total evaporative losses from lakes and reservoirs in the Upper Colorado River Basin were estimated by the U.S. Water Resources Council (1978, p. 17) to be 800,000 acre-ft/yr. In general, evapotranspiration exceeds precipitation from April to October; during the remainder of the year, especially at altitudes above 6,000 ft, evapotranspiration is a minor component of the water budget (see, for example, Waddell and others, 1981, p. 7; Geldon, 1985a, p. 14).

DRAINAGE

The watershed of the Colorado River is the ninth largest river system in the country in terms of discharge and the fifth largest system in terms of drainage area (Iseri and Langbein, 1974). West of the Continental Divide, only the Columbia and Sacramento Rivers have larger discharges; only the Columbia River has a larger drainage area.

The Colorado River is formed by the confluence of five major tributaries—the Green, Upper Colorado, San Juan, Little Colorado, and Gila Rivers. All but the Little Colorado and Gila join the Colorado in the upper part of the basin, which was formally defined by the Colorado River Compact of 1922. This compact officially divided the upper and lower parts of the Colorado River watershed at a point 1 mi downstream from the confluence of the Paria and Colorado Rivers near Lees Ferry, Ariz. This point has no hydrological significance in terms of either the surface-water or ground-water systems.

Prior to 1921, the Colorado River above the confluence with the Green River was known as the Grand River, and the combined flows of the Green and Grand Rivers were known as the Colorado River. On July 25, 1921, the Colorado Legislature extended the name "Colorado River" to the headwaters of the Grand River for political reasons (Fradkin, 1981, p. 35).

The Little Colorado and Gila Rivers, formerly integral parts of the Colorado River system, are now depleted severely by irrigation diversions and are mostly dry channels except during seasonally high flows. The Colorado River itself is depleted severely by irrigation demands in its lower reaches and no longer flows to its historical outlet, the Gulf of California (Fradkin, 1981, p. 319–341).

In the upper basin, the Green and Colorado Rivers and their tributaries (fig. 12) comprise more than 17,000 mi of streams with more than 322,000 acres of lakes and reservoirs (U.S. Water Resources Council, 1978, p. 10). The combined storage capacity of the 82 largest reservoirs in the basin (individual storage capacities of more than 5,000 acre-ft) is about 38,000,000 acre-ft (Liebermann and others, 1988, p. 16–18). Lake Powell, with a storage capacity of 27,000,000 acre-ft, contains 71 percent of the surface-water storage. Other reservoirs with a storage capacity of more than 200,000 acre-ft include Flaming Gorge and Fontenelle (Green River), Navajo (San Juan River), Strawberry (Strawberry River), Blue Mesa (Gunnison River), Lake Granby (Colorado River), McPhee (Dolores River), and Dillon (Blue River).

In the words of Powell (1895), the streams of the upper basin,

> Born in the cold and gloomy solitudes of the * * * mountain region, have a strange, eventful history as they pass down through gorges, tumbling in cascades and cataracts, until they reach the hot, arid plains of the Lower Colorado.

The Colorado River begins on the east side of Mount Richthofen, in the Front Range of Colorado (Iorns and others, 1965, p. 2). The Green River originates at Peak Lake, beneath Fremont Peak in the Wind River Mountains of Wyoming (Fradkin, 1981, p. 34–42). The two streams converge below the White Rim Bench, at the head of Cataract Canyon, 60 mi southeast of Green River, Utah, and 35 mi southwest of Moab, Utah. Historically, the Colorado River was constrained in deep canyons—Cataract, Glen, Marble, and Grand—for almost all of the more than 500 river miles between the confluence point and the Grand Wash Cliffs in Arizona. The lower part of Cataract Canyon and most of Glen Canyon are now flooded by Glen Canyon Dam and Lake Powell.

The Colorado River at its confluence with the Green has a larger discharge but a smaller drainage area. This situation results from the distinctly different topography of the two

[1]Includes reports published under the name of predecessors of the National Oceanic and Atmospheric Administration, including the U.S. Weather Bureau, 1930 to 1964, and the Environmental Science Service Administration, 1965 to 1970.

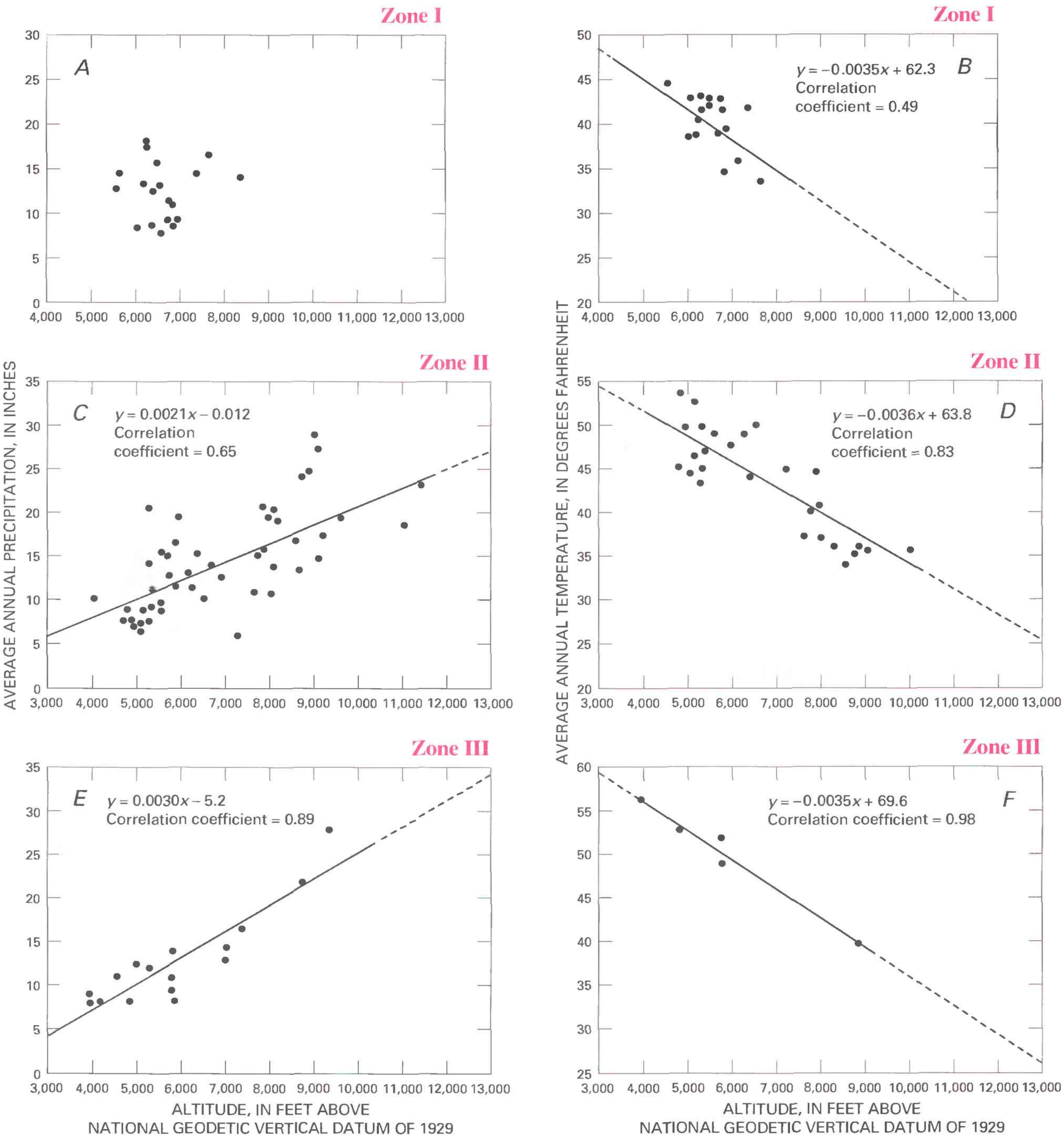


FIGURE 11.—Relations of precipitation and temperature to altitude in the Upper Colorado River Basin. (Plots *I* and *J* modified from Geldon, 1985a, p. 11; plots *K* and *L* modified from Cooley and others, 1969, p. 29.)

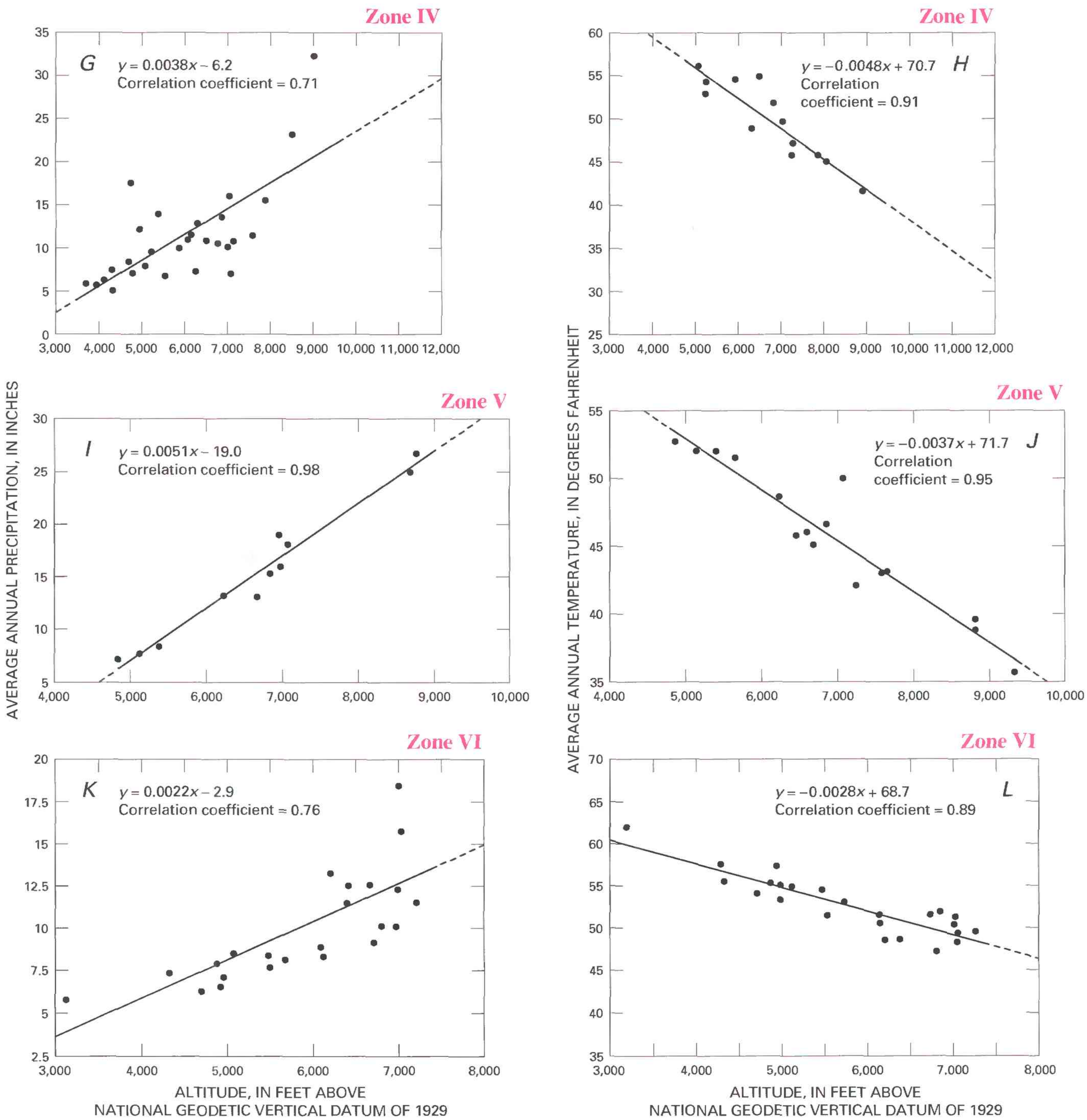


FIGURE 11.—Relations of precipitation and temperature to altitude in the Upper Colorado River Basin. (Plots *I* and *J* modified from Geldon, 1985a, p. 11; plots *K* and *L* modified from Cooley and others, 1969, p. 29)—Continued.

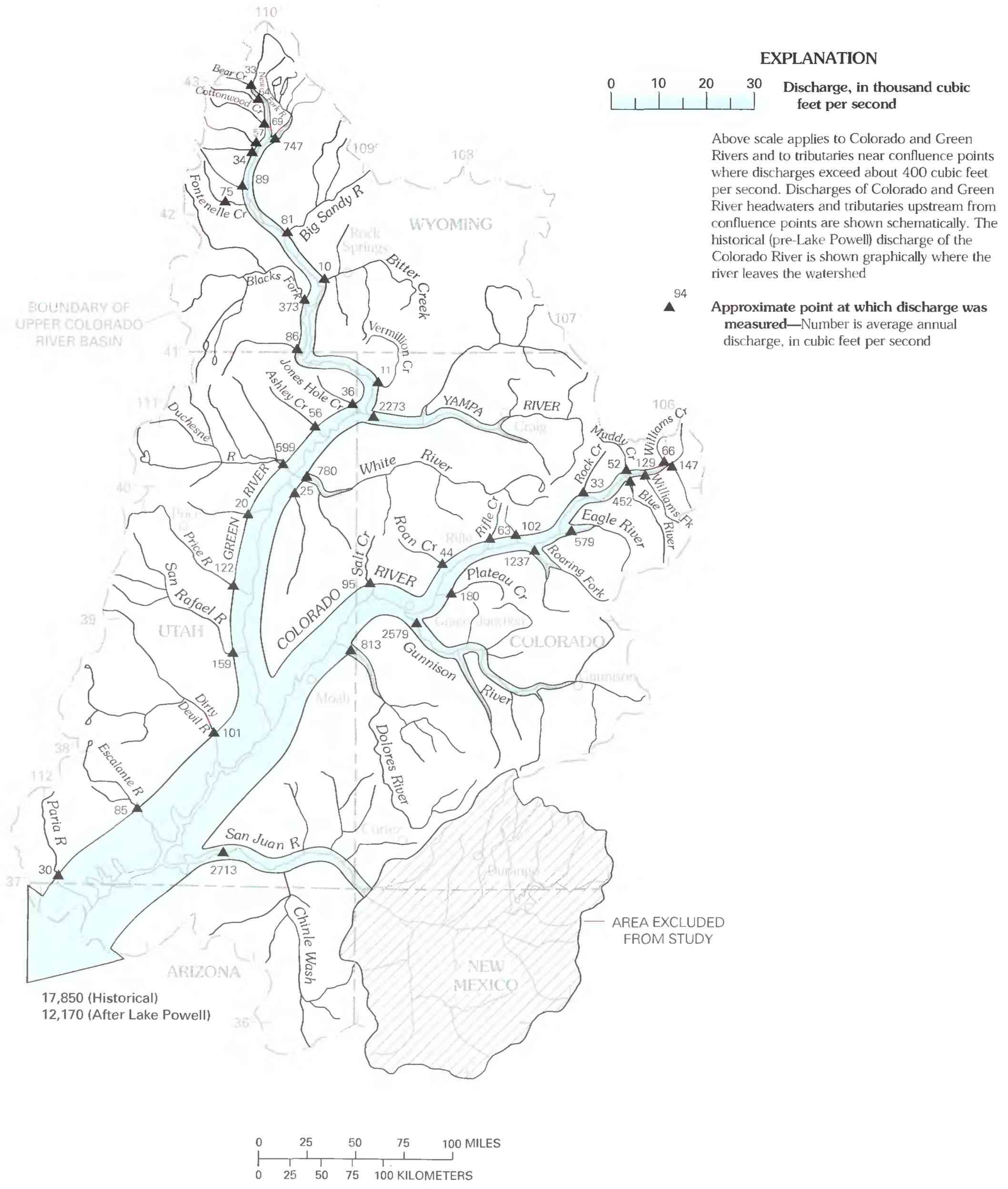


FIGURE 12.—Watersheds and average discharges of the Green and Colorado Rivers upstream from Lees Ferry, Arizona. (Compiled from Druse and Rucker, 1984; Duncan and others, 1984a, 1984b; ReMillard and others, 1984; and White and Garrett, 1982.)

watersheds. Upstream from the confluence, the Colorado River flows mostly through mountainous terrain with a humid climate. However, the Green River flows mostly through arid and semiarid plains and plateaus.

The Colorado River at its confluence with the Green River has an average discharge of about 8,000 ft^3/s. The principal tributaries of the Upper Colorado River (those with individual discharges of more than 150 ft^3/s) are the Gunnison River, Roaring Fork, Dolores River, Eagle River, Blue River, and Plateau Creek (fig. 12). The Green River at its mouth has an average discharge of about 6,600 ft^3/s. Its principal tributaries are the Yampa River, White River, New Fork, Duchesne River, Blacks Fork, San Rafael River, and Price River (fig. 12). Downstream from the confluence of the Colorado and Green Rivers, four rivers—the San Juan, Dirty Devil, Escalante, and Paria—enter the Colorado River (fig. 12). Prior to construction of Glen Canyon Dam and the filling of Lake Powell, the Colorado River at Lees Ferry, Ariz., had an average discharge of 17,850 ft^3/s. After the gates of the dam were closed in 1963, the discharge at Lees Ferry (as of 1982) averaged 12,170 ft^3/s, a net reduction from the preconstruction discharge of 5,680 ft^3/s.

The Colorado and Green Rivers gain streamflow from tributaries, ground-water discharge, and irrigation return flows and are depleted by transbasin diversions, irrigation diversions, seepage to aquifers, and evapotranspiration. On the basis of measured springflows and previous assessments of ground-water contributions by Rush and others (1982), URS Corporation (1983), and Warner and others (1985), cumulative spring and seep inflows to the Green and Colorado Rivers between the most upstream gaging stations and Lees Ferry, Ariz., were determined to be 833 ft^3/s. On the basis of unaccounted gains in these reaches, irrigation return flows are estimated to be at least 739 ft^3/s. Less than 10 percent of the streamflow in the Green and Colorado Rivers, thus, is attributed to direct ground-water discharge and irrigation return flow.

Transbasin diversions occur from the Colorado, Green, Fraser, Blue, Gunnison, Eagle, Fryingpan, Little Snake, Big Sandy, Duchesne, Strawberry, Navajo, Piedra, San Juan and San Rafael Rivers, Los Pinos River, the Williams and Roaring Forks of the Colorado River, and the Henrys Fork of the Green River (Liebermann and others, 1988, p. 14–15). According to Liebermann and others (1988, p. 13), total transbasin diversions from the UCRB during water years 1973–82 averaged 1,020 ft^3/s annually.

Water for irrigation is diverted from the Colorado and Green Rivers and most sizable tributaries. Measured losses from the Green River in Wyoming average 269 ft^3/s. Most of this is attributable to irrigation diversion, but some may be due to evaporation from Fontenelle Reservoir. Irrigation diversions from the Colorado River in Middle Park and Grand Valley, Colo., are estimated to average as much as 353 ft^3/s. If about 65 percent of the water diverted for irrigation in the Upper Colorado River Basin returns to streams, as indicated by the U.S. Water Resources Council (1978, p. 17), then on the basis of return flows reported earlier (739 ft^3/s), total irrigation diversion from the Colorado and Green Rivers could be about 1,140 ft^3/s.

Since 1963, the largest water losses from the Colorado and Green Rivers have been the result of storage in Lake Powell, evaporation from the reservoir, and seepage of water from the reservoir into adjacent Mesozoic and Paleozoic rocks. According to Thomas (1986, p. 14–15), annual leakage from the reservoir to the underlying aquifers is about 600,000 acre-ft/yr. As a result of this leakage, ground-water levels in wells within 10 mi of the reservoir are known or estimated (by finite-difference modeling) to have risen by less than 25 to 500 ft between 1963 and 1983.

Except for regulated reaches downstream from dams, streamflows vary substantially during the year throughout the watershed. Seasonal flows typically are highest in spring and early summer and lowest in late fall and early winter. About 70 percent of the annual runoff occurs from early April to late July (U.S. Water Resources Council, 1978, p. 10). This runoff is derived mostly from snowmelt. Thunderstorms produce high flows of short duration from July through October. From November to April, most streams are at or near base flow, except during periods of storm-induced runoff. As seen in the 1984 records of the Colorado River below Glenwood Springs, Colo., and the Green River near Green River, Wyo. (fig. 13), monthly discharges of perennial streams in mountainous regions are similar in timing to, but larger in amplitude (per square mile of drainage area) than, these fluctuations in lowland regions. As seen in the 1982 record of the Colorado River at Lees Ferry, Ariz. (fig. 13), seasonal discharge fluctuations are suppressed in reaches regulated by dams; such discharge fluctuations are controlled mainly by the timing of demands for water.

REGIONAL GEOLOGY

Rocks exposed in the UCRB are Precambrian to Tertiary in age (fig. 14). The Precambrian and Paleozoic rocks typically are exposed in the center and on the flanks of uplifted areas. Mesozoic rocks have widespread surface exposure, particularly in the southern part of the UCRB. Tertiary sedimentary rocks form most of the surface area in the Piceance, Uinta, Sand Wash, and greater Green River Basins. Tertiary intrusive and extrusive rocks are widespread in the Colorado Plateaus and Southern Rocky Mountains physiographic provinces. Thrust faults, particularly in the Overthrust Belt, Gros Ventre Range, Wind River Mountains, and Uinta Mountains, and high-angle normal and reverse faults in the High Plateaus region, Uncompahgre Plateau, Paradox Basin, and the Southern Rocky Mountains have had a pronounced effect on the distribution of the exposed rocks.

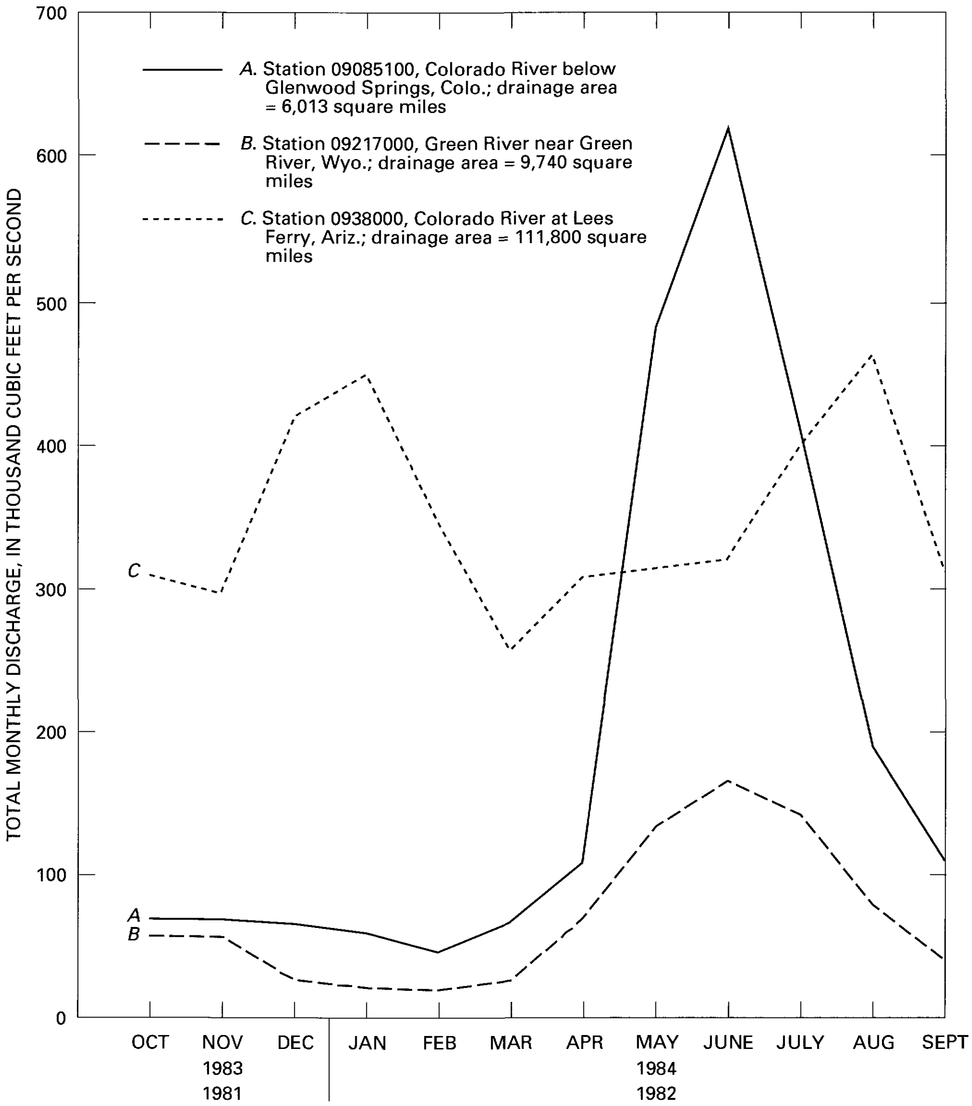


FIGURE 13.—Total monthly discharges of the Colorado River below Glenwood Springs, Colorado, and the Green River near Green River, Wyoming, during water year 1984, and for the Colorado River at Lees Ferry, Arizona, during water year 1982. (Water year 1984 discharge for the Colorado River at Lees Ferry is not shown because of atypically large releases from Glen Canyon Dam in response to floodwaters entering Lake Powell during that year.)

Precambrian granitic, metamorphic, and sedimentary rocks underlie the Paleozoic rocks (fig. 15) with subtle to pronounced angular unconformity. Archean to Late Proterozoic granitic and metamorphic rocks compose most of the Precambrian complex. However, slightly metamorphosed to unmetamorphosed sedimentary and volcanic rocks are present in a few areas. In the San Juan Mountains and vicinity, an 8,000-ft-thick sequence of quartzite, slate, and phyllite, belonging to the Early and Middle Proterozoic Uncompahgre Formation, underlies the Paleozoic rocks. In and near the Uinta Mountains and Uinta Basin, a 24,000-ft-thick sequence of quartzite, sandstone, and shale, belonging to the Middle and Late Proterozoic Uinta Mountain Group, underlies the Paleozoic rocks. From the Grand Canyon to the vicinity of the Paria River in northeastern Arizona and southeastern Utah, a 13,000-ft-thick sequence of quartzite, sandstone, conglomerate, breccia, shale, limestone, and basalt, belonging to the Middle to Late Proterozoic Unkar Group, Nankoweap Formation, and Chuar Group, underlies the Paleozoic rocks.

The thickness of Paleozoic formations present in the UCRB ranges from 0 to about 18,000 ft (fig. 16). All systems except the Silurian are represented. The Paleozoic rocks were deposited during five advances and retreats of the sea. Cambrian to Lower Mississippian formations consist mostly of limestone and dolomite, although thick intervals of sandstone, quartzite, and shale

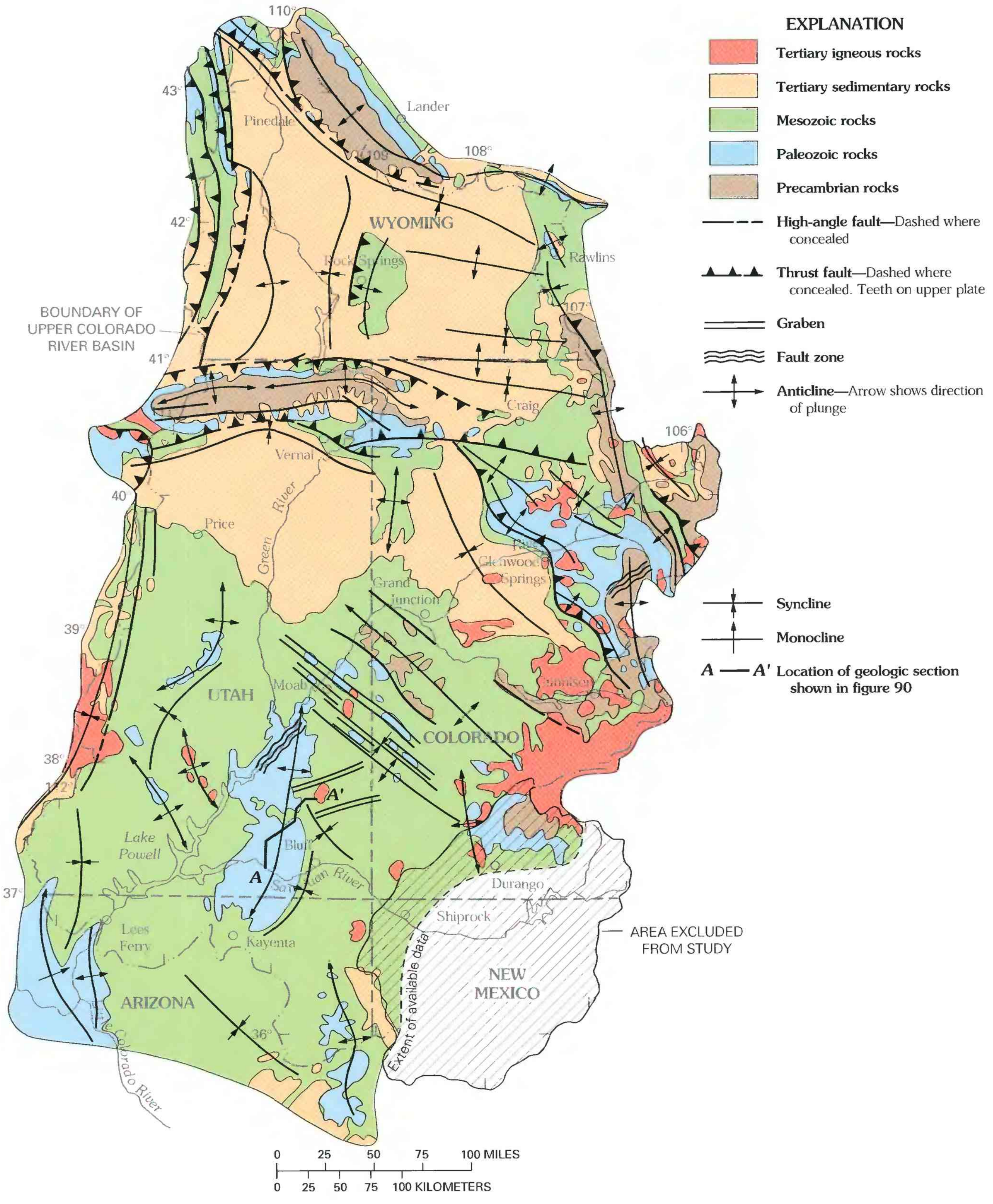


FIGURE 14.—Generalized geology of the Upper Colorado River Basin and vicinity. (Modified from Geldon, in press, pls. 1 and 2.)

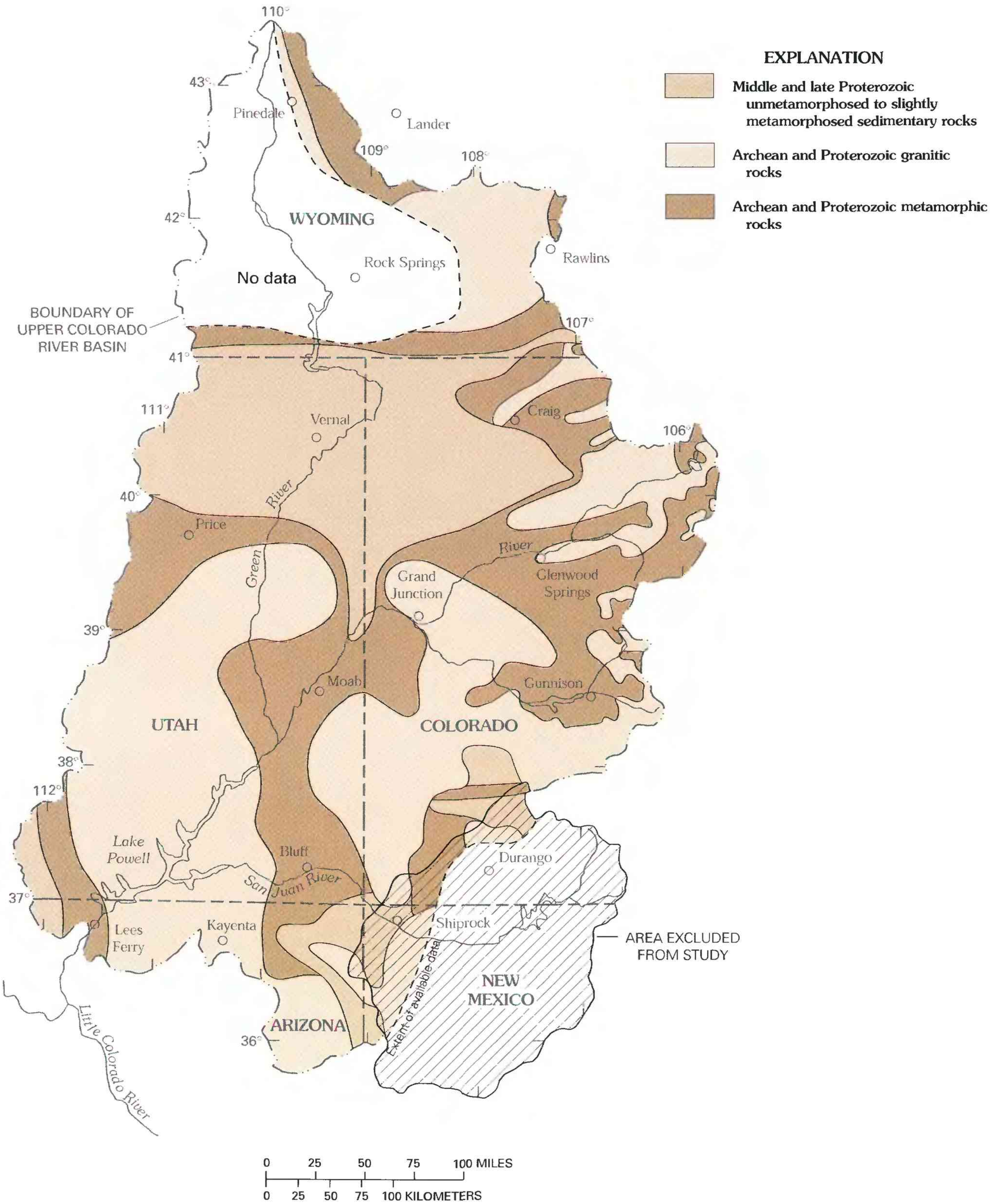


FIGURE 15.—Distribution of Precambrian rocks in the Upper Colorado River Basin. (Modified from Geldon, in press, pl. 4.)

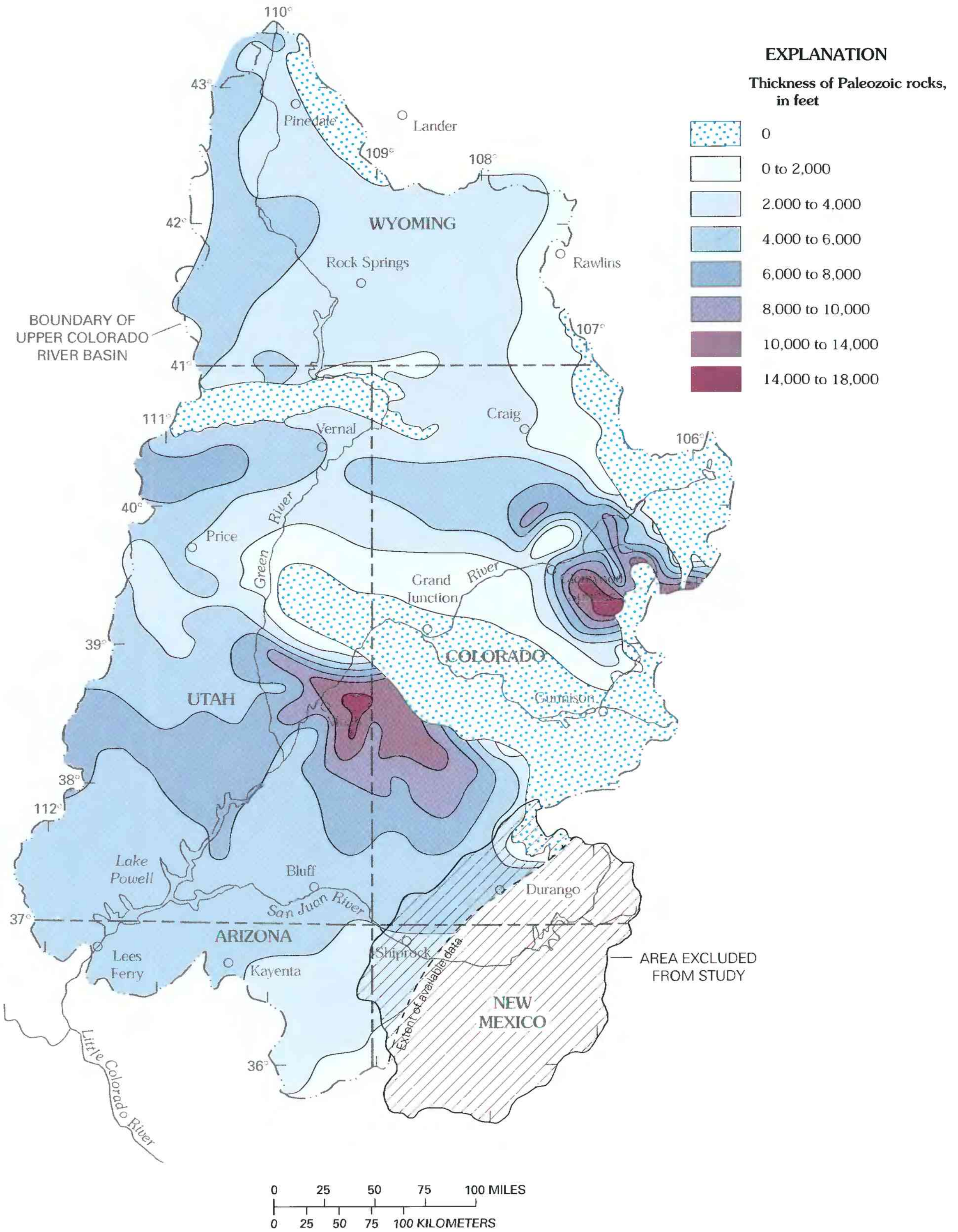


FIGURE 16.—Distribution and thickness of the Paleozoic rocks in the Upper Colorado River Basin. (Modified from Geldon, in press, pl. 3.)

are present. These lower and middle Paleozoic formations are overlain by much thicker Upper Mississippian to Permian formations. The younger formations consist mostly of sandstone, conglomerate, and shale, but variably thick intervals of carbonate rocks occur, and thick deposits of bedded to diapiric anhydrite and halite are present, particularly in the Paradox and Eagle Basins. Deposition of the Upper Mississippian to Permian formations was affected substantially by uplift and subsidence. Consequently, pronounced lateral variations in lithology are characteristic of the Pennsylvanian and Permian rocks.

The Paleozoic rocks are overlain in most areas by Triassic formations that are predominantly composed of shale. Locally, lower parts of the Triassic formations in contact with the Paleozoic rocks consist partly or entirely of sandstone, conglomerate, or limestone (fig. 17). Contacts between the Paleozoic and Triassic rocks are gradational to unconformable. In small areas within the Southern Rocky Mountains and Piceance Basin, the Triassic rocks were removed by pre-Jurassic erosion, and the Wingate and Entrada Sandstones of Jurassic age overlie the Paleozoic rocks with pronounced angular unconformity. Triassic rocks also are missing in a few small areas within the Overthrust Belt. In these areas, the Wasatch Formation, a unit of Tertiary age consisting mostly of shale and fine-grained sandstone, unconformably rests upon the Paleozoic rocks (fig. 17). The total thickness of Mesozoic and Tertiary rocks overlying the Paleozoic rocks ranges from 0 to about 27,000 ft (fig. 18).

GENERAL HYDROLOGIC PROPERTIES OF THE PALEOZOIC ROCKS

The 11 hydrogeologic units consisting of Paleozoic rocks each possess distinctive lithologic and hydrologic properties, which are summarized in table 5. For comparison, typical ranges in permeability and hydraulic conductivity exhibited by sedimentary rocks are shown in figure 19. Use of the relative terms small, moderate, and large in discussions of permeability and hydraulic conductivity in this report are referenced to this figure. In regard to transmissivity and yield, the following ranges apply:

Relative term	Transmissivity (feet squared per day)	Yield (gallons per minute)
Small	0.0005–1	<1–50
Moderate	1–100	50–500
Large	100–50,000	500–50,000

Indicative of variable lithology and tectonic setting, all hydrogeologic units exhibit ranges in hydrologic properties spanning several orders of magnitude (table 5). Variations within a hydrogeologic unit vertically can be as large as variations laterally across the UCRB. Ranges overlap but, in general, aquifers are characterized by larger values of unit-averaged porosity, permeability, and hydraulic conductivity, composite transmissivity, and sustained yields than confining units.

All hydrogeologic units are most permeable and, therefore, have the largest yields of water in uplifted areas. Fractures associated with uplift serve as conduits for downward percolation of meteoric water and streamflow and connect confining units with aquifers, resulting in a localized capability of confining units to supply water. Conversely, all hydrogeologic units decrease in permeability and water-supply capability away from uplifted areas (toward structural basins) as fractures associated with uplift become less common. The permeability and water-supply capabilities of hydrogeologic units in structural basins have been reduced further by compaction beneath thousands of feet of Mesozoic and Tertiary rocks and cementation by mineral-laden solutions migrating down structural and topographic gradients.

Among rock types commonly present in the hydrogeologic units composed of Paleozoic rocks, sandstone, with a median porosity of 8.0 percent, is the most porous (fig. 20). Next, in order of decreasing median porosity, are dolomite (4.6 percent), limestone (2.8 percent), shale (2.4 percent), and anhydrite and halite (0.4 percent). Among clastic rocks, porosity is smallest in sandstone cemented with silica, calcite, dolomite, or anhydrite; intermediate in sandstone with matrix minerals, such as glauconite or clay; and largest in generally friable, micaceous and arkosic sandstone (fig. 21). The porosity of clastic rocks also increases as the grain size increases and is larger in coarse-grained and gravelly sandstone than in shale or very fine-, fine-, or medium-grained sandstone.

Among carbonate rocks, porosity generally is smallest in varieties containing impurities that impart a laminar texture, such as shaly, carbonaceous, or algal limestone and dolomite, and largest in varieties that are susceptible to dissolution around inclusions, such as cherty, anhydritic, or vuggy limestone and dolomite. The porosity of carbonate rocks also varies with increasing graininess and is larger in chalky, earthy, sandy, and sucrosic varieties than in crystalline and cryptocrystalline (fine-grained) varieties. The porosity of brecciated limestone and dolomite appears to be small, despite an inherently fractured nature, and probably was reduced by calcite cement in the fractures of the samples that were analyzed.

Pore-scale permeability, like porosity, is influenced mainly by sedimentary composition and texture. The median pore-scale permeability of sandstone in this study was 4 times larger than the median pore-scale permeability of dolomite and 10 times larger than the median pore-scale permeability of limestone

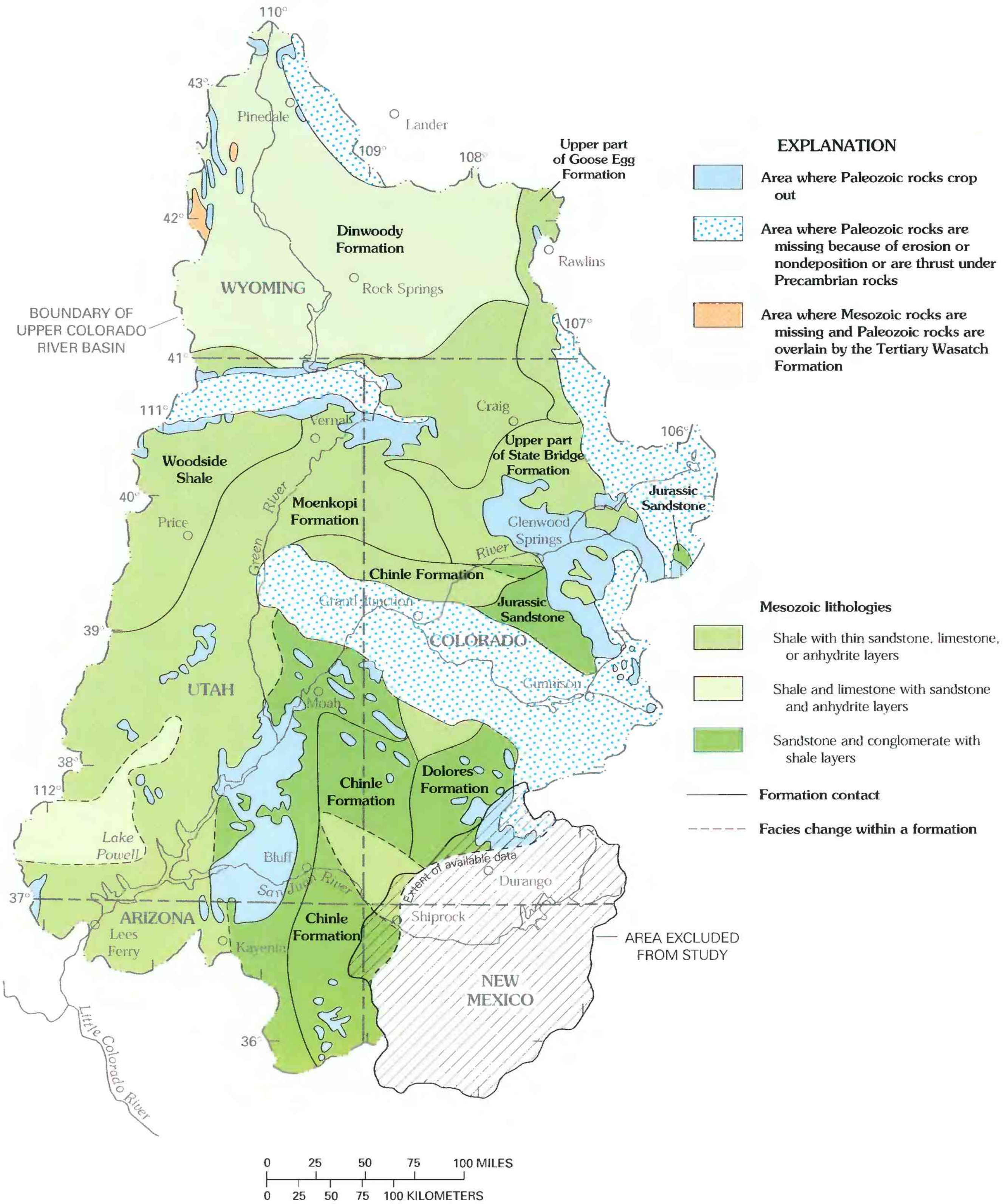


FIGURE 17.—Lithology of the Mesozoic rocks in contact with Paleozoic rocks in the Upper Colorado River Basin. (Modified from Geldon, in press, pl. 4.)

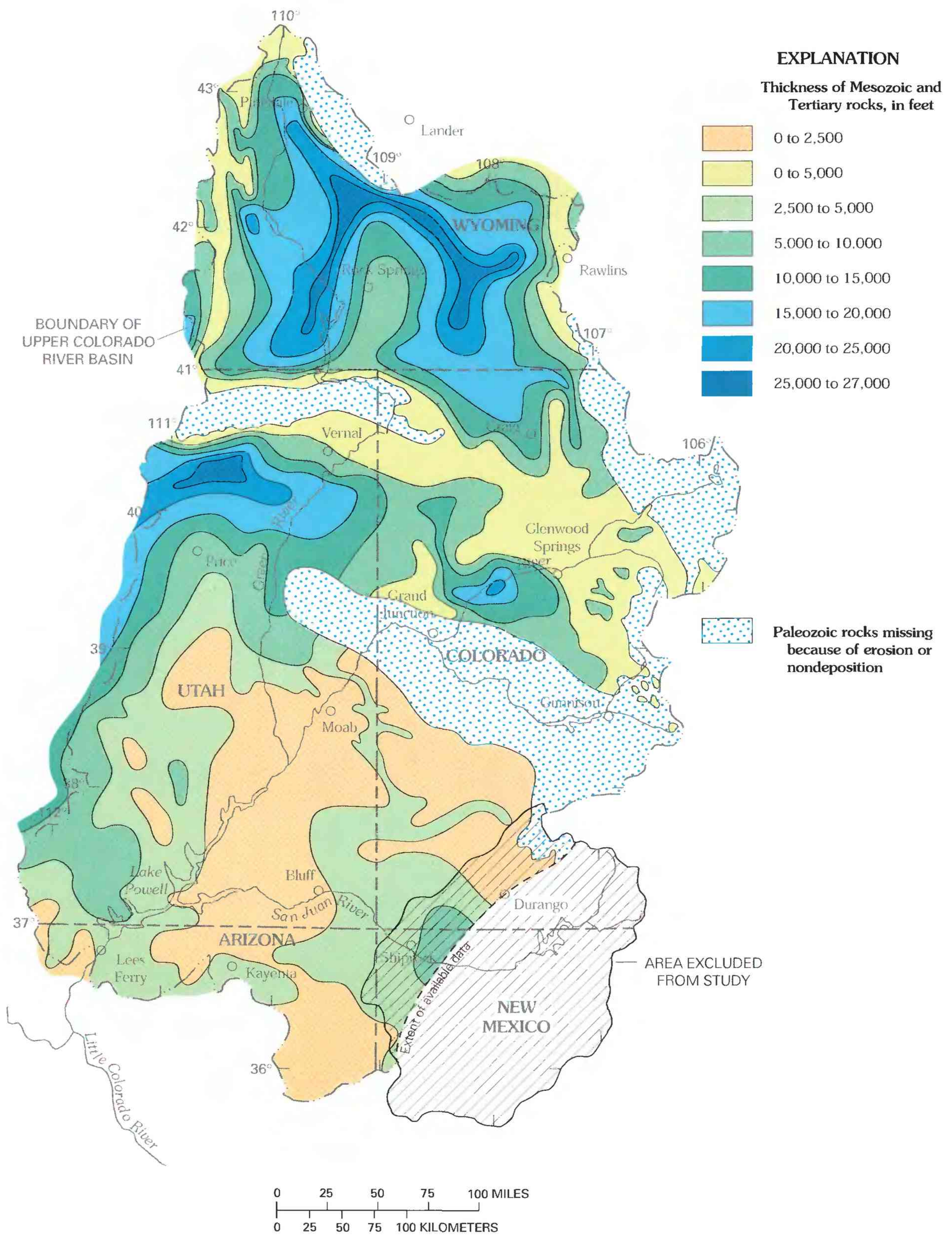


FIGURE 18.—Thickness of the Mesozoic and Tertiary rocks above the Paleozoic rocks in the Upper Colorado River Basin. (Modified from Geldon, in press, pl. 3.)

TABLE 5.—*Summary of lithologic and hydrologic properties of the hydrogeologic units composed of Paleozoic rocks in the Upper Colorado River Basin*

[<, less than]

	Hydrogeologic unit	General lithology	Thickness (feet)	Porosity (percent)	Permeability (millidarcies)	Hydraulic conductivity (feet per day)	Transmissivity (feet squared per day)	Artesian yields from wells and springs (gallons per minute)
Canyonlands aquifer	Park City–State Bridge zone	Variable proportions of shale, sandstone, carbonate rocks, gypsum-anhydrite, chert, and phosphorite.	0–800	Range <0.1 to 22. Median values are: sandstone - 7.4; limestone - 3.4; dolomite - 2.1. Unit-averaged range is 2 to 6.	Pore-scale range is <0.01 to 1,450. Local-scale range is 0.0049 to 330. Medians of pore-scale values are: dolomite - 0.06; sandstone - 0.04; limestone - 0.01.	Range is 0.000012 to 0.80, with a median of 0.006. Unit-averaged range is 0.00001 to 1.	Composite values range from 0.005 to 70, depending on lithology and development of fracture and solution channels. Largest values in uplifted areas.	Typical range is <1 to 700. Kendall Warm Spring discharges 2,870 in association with a thrust fault.
	Weber-De Chelly zone	Quartz sandstone with minor conglomerate, shale, and carbonate rocks.	0–4,000	Range <0.1 to 28. Median value for sandstone is 8.0. Unit-averaged range is 1 to 28.	Strong correlation between porosity and pore-scale permeability for sandstone. Range of pore-scale values is <0.01 to 3,080. Range of local-scale values is 0.014 to 380. Median pore-scale sandstone permeability value is 0.40.	Range is 0.000034 to 61, with a median of 0.012. Unit-averaged range is 0.00005 to 20.	Composite values range from 0.01 to 6,000, depending on cementation and to a lesser degree, on fracturing. Values decrease from uplifted areas to basins.	Typical range is <1 to 2,200. Three springs downstream from sinks in Brush Creek, Ashley Creek, and Dry Fork Ashley Creek have discharges of 36,000 to 90,000 during peak runoff.
	Cutler-Maroon zone	Variable proportions of arkosic to quartzose sandstone, conglomerate, shale, and carbonate rocks.	0–16,500	Range 0.4 to 20. Median values are: arkosic sandstone - 11.6; quartz sandstone - 7.3; limestone - 2.4. Unit-averaged range is 2 to 14.	Pore-scale range is 0.00071 to 232. Local-scale range is 0.00078 to 68. Medians of pore-scale values are: sandstone - 0.46; limestone - 0.01	Range is 0.000002 to 10, with a median of 0.15. Unit-averaged range is 0.00005 to 10.	Composite values range from 0.0005 to 10,000, depending on thickness, lithology, and development of fractures and solution channels. Largest values in uplifted areas.	Typical range is <1 to 900. Jones Hole Springs, which occur at a subregional discharge point for the entire aquifer system, have a combined discharge of 16,550.
Four Corners confining unit	Paradox–Eagle Valley subunit	Variable proportions of shale, gypsum-anhydrite, halite, limestone, and dolomite, with subordinate sandstone.	0–9,700 where bedded: salt diapirs as much as 15,000	Range <0.1 to 32. Median values are: sandstone - 6.6; dolomite - 6.1; limestone - 3.5; shale - 1.8; anhydrite - 0.4; halite - 0.4. Unit-averaged range is <1 to 6.	Pore-scale range is <0.0001 to 3,460. Local-scale range is 0.0031 to 550. Medians of pore-scale values are: dolomite - 0.21; limestone - 0.10; anhydrite and halite - <0.01; sandstone - 0.01; shale - <0.01;	Range is 0.000008 to 1.5, with a median of 0.005. Unit-averaged range is 0.00001 to 0.03.	Transmissivity probably depends on lithology and the development of fractures and solution channels. Based on lithology alone, composite transmissivity is estimated to range from 0.05 to 15.	Typical range is <1 to 150. Big Spring at Dotsero, Colo., discharges 450 in association with faults.
	Belden-Molas subunit	Shale with subordinate carbonate rocks and minor gypsum-anhydrite and sandstone.	0–4,300	Range and median values for individual rock types probably similar to above unit. Unit-averaged range is <2 to 6.	Range in pore-scale permeability and median values for individual rock types probably similar to above unit. Range in local-scale permeability is 0.0012 to 130.	Range is 0.000013 to 0.54, with a median of 0.0046. Unit-averaged range is 0.00001 to 0.10.	Composite values range from 0.001 to 50, depending mainly on development of fractures and solution channels, which decrease away from uplifted areas.	Typical range is <1 to 100. A spring northwest of Glenwood Springs, Colo., has a discharge of 266.
Four Corners aquifer system	Madison aquifer-Darwin-Humbug zone	Interbedded sandstone, shale, carbonate rocks, and breccia, with minor gypsum.	0–800	Range 1.4 to 16 in 13 samples of sandstone.	Pore-scale permeability of sandstone ranged from 0.08 to 60 in 13 samples. Local-scale permeability of dolomite, sandstone, and shale in a drill-stem test near Maybell, Colo., was 292.	0.71 in Maybell drill-stem test.	Composite transmissivity estimated to be 110 at site of Maybell drill-stem test, based on thickness and hydraulic conductivity; could be as much 1,000 where aquifer consists mostly of fractured and cavernous limestone.	7.2 to 34 in four drill-stem tests. However, a spring in the Uinta Mountains discharges 900.
	Madison aquifer-Redwall-Leadville zone	Limestone and dolomite with minor bedded and nodular chert, shale, and sandstone. Increasing dolomite in proportion to limestone from uplifts to basins.	0–2,500	Range 0.3 to 22. Median values are: dolomite - 6.5; limestone - 1.7. Unit-averaged range is <1 to 11.	No correlation between porosity and permeability. Pore-scale range is <0.01 to 940. Local-scale range is 0.02 to 1,800. Medians of pore-scale values are: dolomite - 0.10; limestone - <0.01.	Range is 0.00005 to 200, with a median of 0.023. Unit-averaged range is 0.00005 to 200.	Composite values range from 0.01 to 47,000, primarily depending on development of fractures and solution channels, which decrease away from uplifted areas.	Typical range is <1 to 10,000. Discharges from Blue Spring and nearby springs in the canyon of the Little Colorado River, a subregional discharge area, atypically range from 11,000 to 45,000.

TABLE 5.—*Summary of lithologic and hydrologic properties of the hydrogeologic units composed of Paleozoic rocks in the Upper Colorado River Basin—Continued*

[<, less than]

	Hydrogeologic unit	General lithology	Thickness (feet)	Porosity (percent)	Permeability (millidarcies)	Hydraulic conductivity (feet per day)	Transmissivity (feet squared per day)	Artesian yields from wells and springs (gallons per minute)
Four Corners aquifer system	Elbert-Parting confining unit	Variable proportions of shale, sandstone, quartzite, limestone, dolomite, and gypsum-anhydrite.	0–700	Range in the Four Corners area is 0.3 to 12. Median values are: sandstone - 6.1; dolomite - 1.9. Unit-averaged range is <2 to 5.	For the Four corners area, pore-scale range is <0.01 to 179, and local-scale range is 0.021 to 95. Medians of pore-scale values are: sandstone - 0.42; dolomite - <0.01.	Range in the Four Corners area is 0.000051 to 0.25, with a median of 0.004. Unit-averaged range is 0.00005 to 0.25.	Composite values in the Four Corners area range from 0.008 to 200, depending mainly on lithology and the degree of fracturing.	Typical range is <1 to 60, but two springs near thrust faults have discharges of 900 and 1,100.
	Bighorn aquifer	Limestone and dolomite with subordinate shale and minor sandstone.	0–3,000	Based on laboratory and geophysical measurements at four sites, known range is 0.1 to 13. Unit-averaged range is 1 to 5.	Based on 56 measurements, known range in pore-scale permeability is <0.01 to 157. Local-scale permeability at 13 sites ranges from 0.029 to 23.	Conversion in in-situ permeability data indicates a range of 0.000071 to 0.056, with a median of 0.007.	Composite values based on thickness and hydraulic conductivity at 11 sites range from 0.1 to 6.	Typical range is <1 to 500, but one spring near a thrust fault has a discharge 3,200.
	Gros Ventre confining unit	Shale with subordinate sandstone and carbonate rocks.	0–1,100	No data.	Based on one drill-stem test in this unit and three injection tests in Proterozoic shale, estimated range in local-scale permeability is <0.18 to 29.	Conversion of in-situ permeability data indicates a range of 0.00044 to 0.07.	No data.	Known range from one well and four springs is <1 to 8. A fault-controlled spring has a discharge of 900.
	Flathead aquifer	Sandstone and quartzite with subordinate carbonate rocks, shale, and conglomerate.	0–800	In one borehole, geophysically determined range is 0.5 to 24; median is 12.	Based on six drill-stem tests, known range in local-scale permeability is 0.003 to 42.	Conversion of local-scale permeability data and two aquifer tests of Precambrian quartzite indicate a range of 0.000007 to 3.	Based on six drill-stem tests in the UCRB and three flowing-well tests in the Bighorn Basin, range in composite transmissivity probably is 0.001 to 300, depending on degree of fracturing and cementation.	Range in UCRB and Grand Canyon is <1 to 250, but flowing wells in Bighorn Basin have discharges of 800 to 3,000.

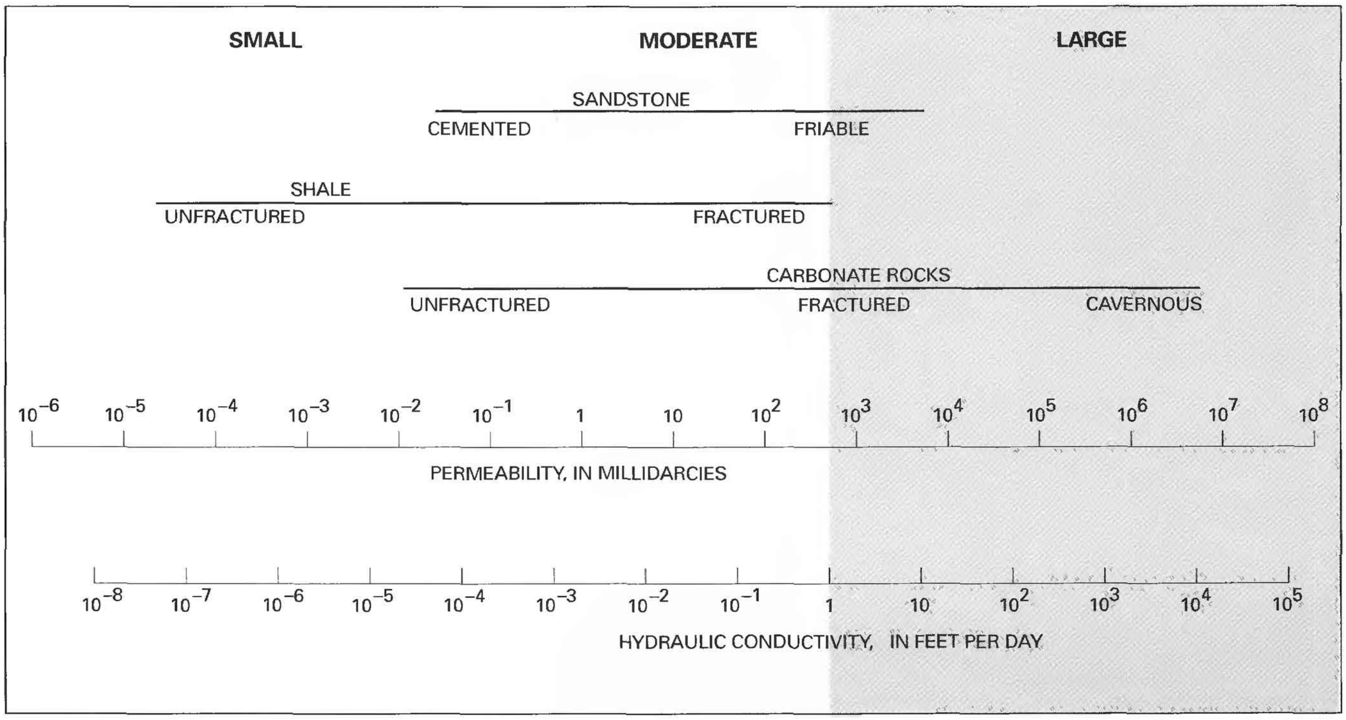


FIGURE 19.—Range in typical permeability and hydraulic-conductivity values of sedimentary rocks. (Modified from Freeze and Cherry, 1979, p. 29; Heath, 1983, p. 13.)

(fig. 22). Median values of pore-scale permeability for shale, anhydrite, and halite were found to be less than 0.01 millidarcy (md), and a few values in the range of 0.0001 to 0.01 md were obtained for each of these rock types. The number of evaporite analyses may be insufficient to be representative of evaporite permeability regionwide. By way of comparison, Thayer (1983, p. 7) reported that the pore-scale permeability of four samples of Mississippian anhydrite in and near the Powder River Basin of Montana and Wyoming ranged from 0.01 to 0.06 md, with a mean value of 0.03 md. Kreitler and others (1985, p. 151) reported that bedded halite in the Salado Formation of Permian age at Carlsbad, N. Mex. (location of city shown on pl. 1) was found to have pore-scale permeability ranging from 0.0001 to 0.001 md and local-scale permeability ranging from 0.003 to 0.02 md.

Sandstone and carbonate textures affect pore-scale permeability in much the same manner as they affect porosity (fig. 23), although some varieties of sandstone, limestone, and dolomite were found to be more or less permeable than otherwise would be expected from their relative porosity. Consistent with results reported by McComas (1963) for the Paradox Basin, algal limestone samples were found to have disproportionately large permeability with respect to porosity. McComas attributed this situation to inherent vugginess in algal reef deposits. Several varieties of limestone, including anhydritic, cherty, shaly, sucrose, and chalky varieties, were found to be more permeable than corresponding varieties of dolomite with larger porosity. This may reflect a tendency reported by McComas of dolomite formed by replacement of limestone to have many unconnected vugs causing small permeability despite large porosity.

Because of lithologic variability, carbonate rocks and sandstone within hydrogeologic units display little correlation between pore-scale permeability and sampling depth. In the Redwall-Leadville zone of the Madison aquifer, for example, the pore-scale permeability of dolomite has the same range at nearly every depth sampled (fig. 24). Similarly, in the Weber-De Chelly zone of the Canyonlands aquifer, the range in the pore-scale permeability of sandstone is virtually independent of sampling depth. However, in the latter hydrogeologic unit, median values of pore-scale permeability decrease with proximity to land surface. These observations imply that the pore-scale permeability of sandstone aquifers probably is affected by leaching of cementing minerals from outcrops and near-surface horizons and redeposition of these minerals in deeper horizons. In carbonate aquifers, pore-scale permeability may be independent of depth because of processes that occurred in the distant past, prior to the rocks being buried to present depths. Such processes could include: (1) Diagenetic alteration of limestone to dolomite by seawater shortly after deposition; and (2) development of karst topography during times when rocks that are now deeply buried were at or near the surface (particularly during the Pennsylvanian-Permian uplift of the ancestral Rocky Mountains).

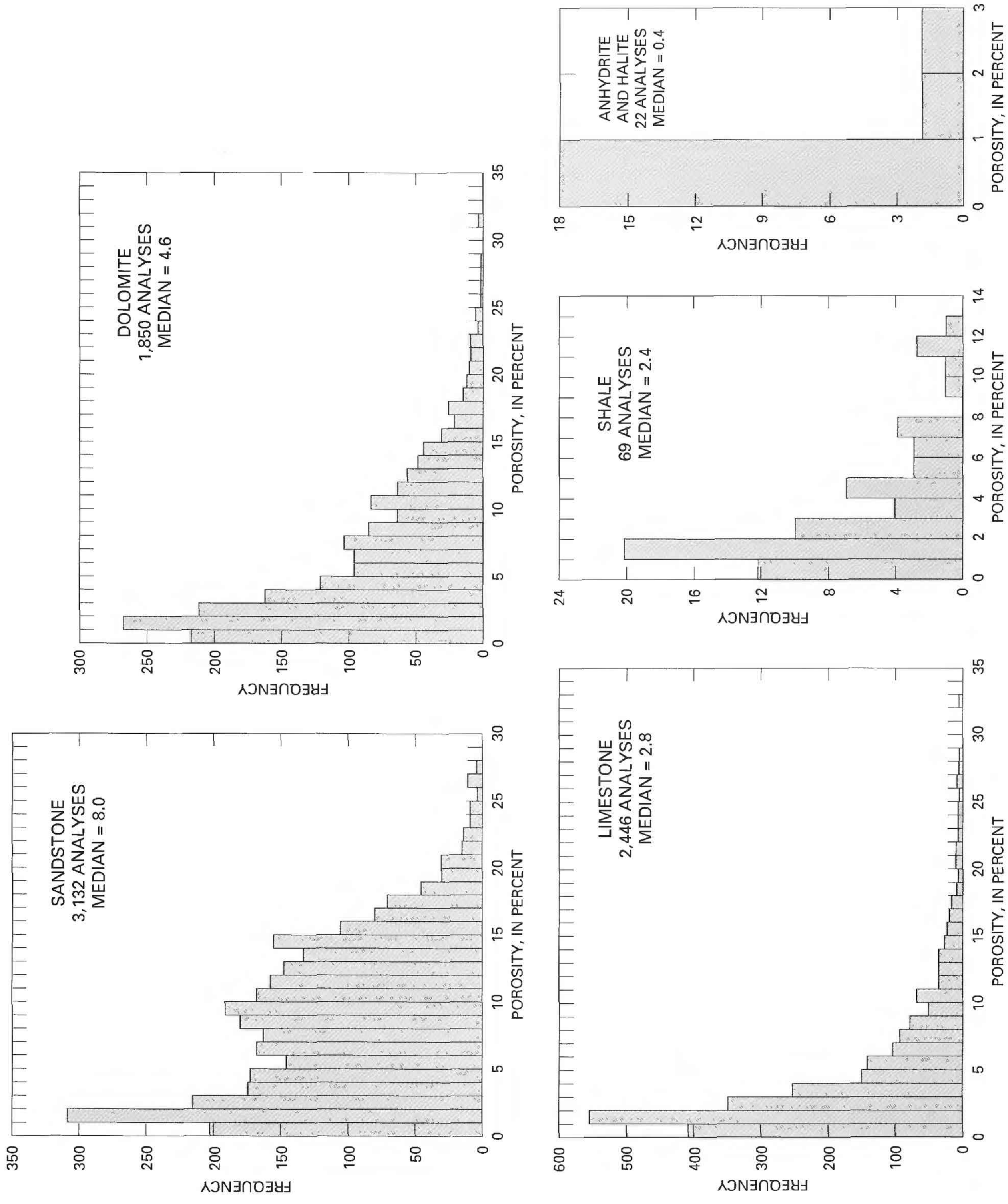


FIGURE 20.—Frequency distribution of porosity in Paleozoic sedimentary rocks of the Upper Colorado River Basin.

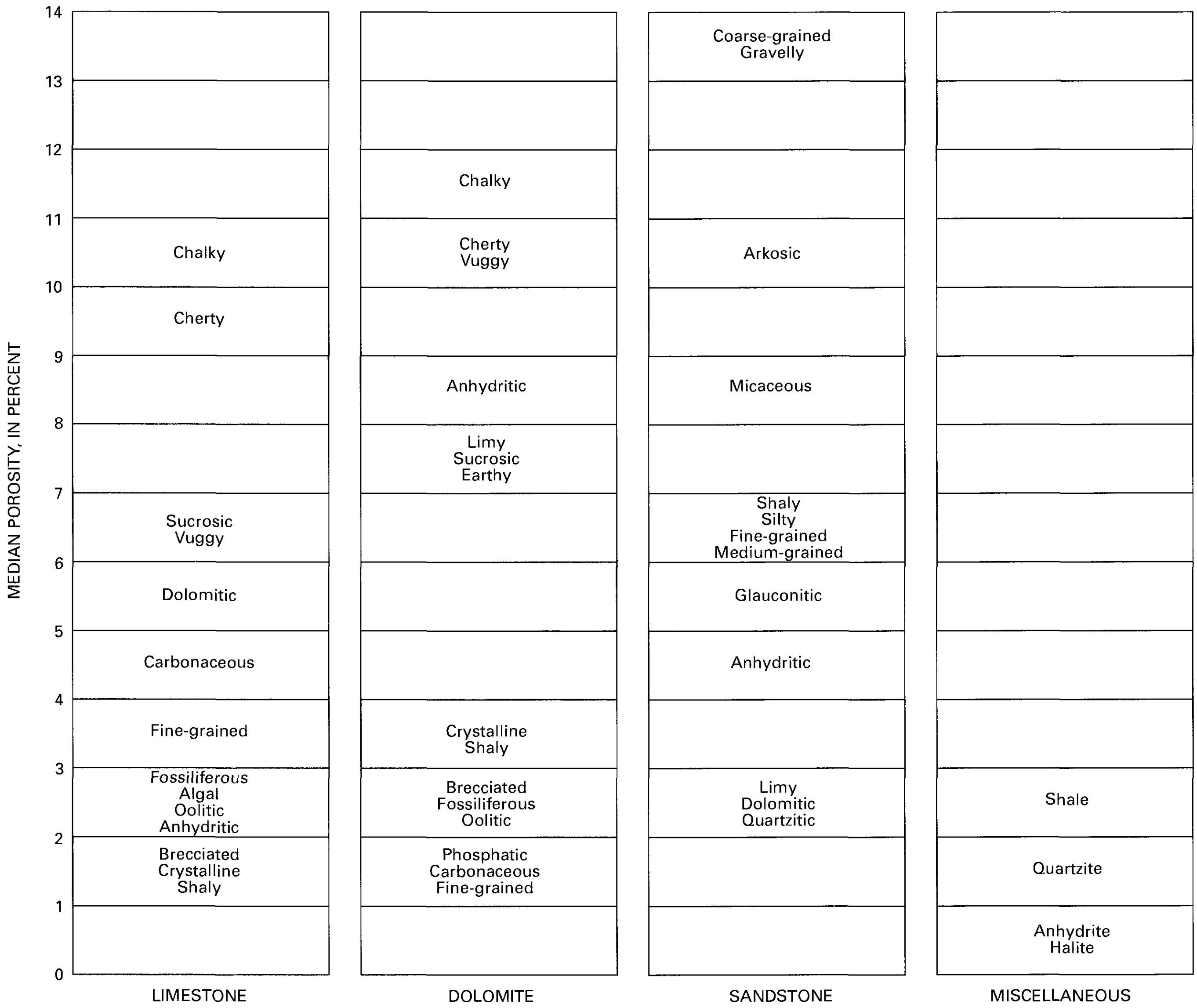


FIGURE 21.—Relation of porosity to variations in the composition and texture of Paleozoic sedimentary rocks of the Upper Colorado River Basin.

HYDROLOGIC PROPERTIES OF THE FOUR CORNERS AQUIFER SYSTEM.

According to Laney and Davidson (1986, p. 6), an aquifer system is defined as

> *** a heterogeneous body of intercalated permeable and poorly permeable material that functions regionally as a water yielding hydraulic unit; it comprises two or more permeable beds (aquifers) separated at least locally by aquitards (confining units) that impede ground-water movement but do not greatly affect the regional hydraulic continuity of the system.

Within the UCRB, the Paleozoic rocks from the top of the Precambrian rocks that form the basal confining unit to the top of the Darwin-Humbug zone of the Madison aquifer that lie immediately below the Four Corners confining unit satisfy the criteria of Laney and Davidson (1986) to be considered a single aquifer system because: (1) Confining units within the interval, although regionally extensive, generally are much thinner than aquifers; (2) in many places, confining units within the interval are no less permeable than aquifers and transmit water with the aquifers as a single entity; and (3) confining units within the interval pinch out locally, such that overlying and underlying aquifers are in direct contact.

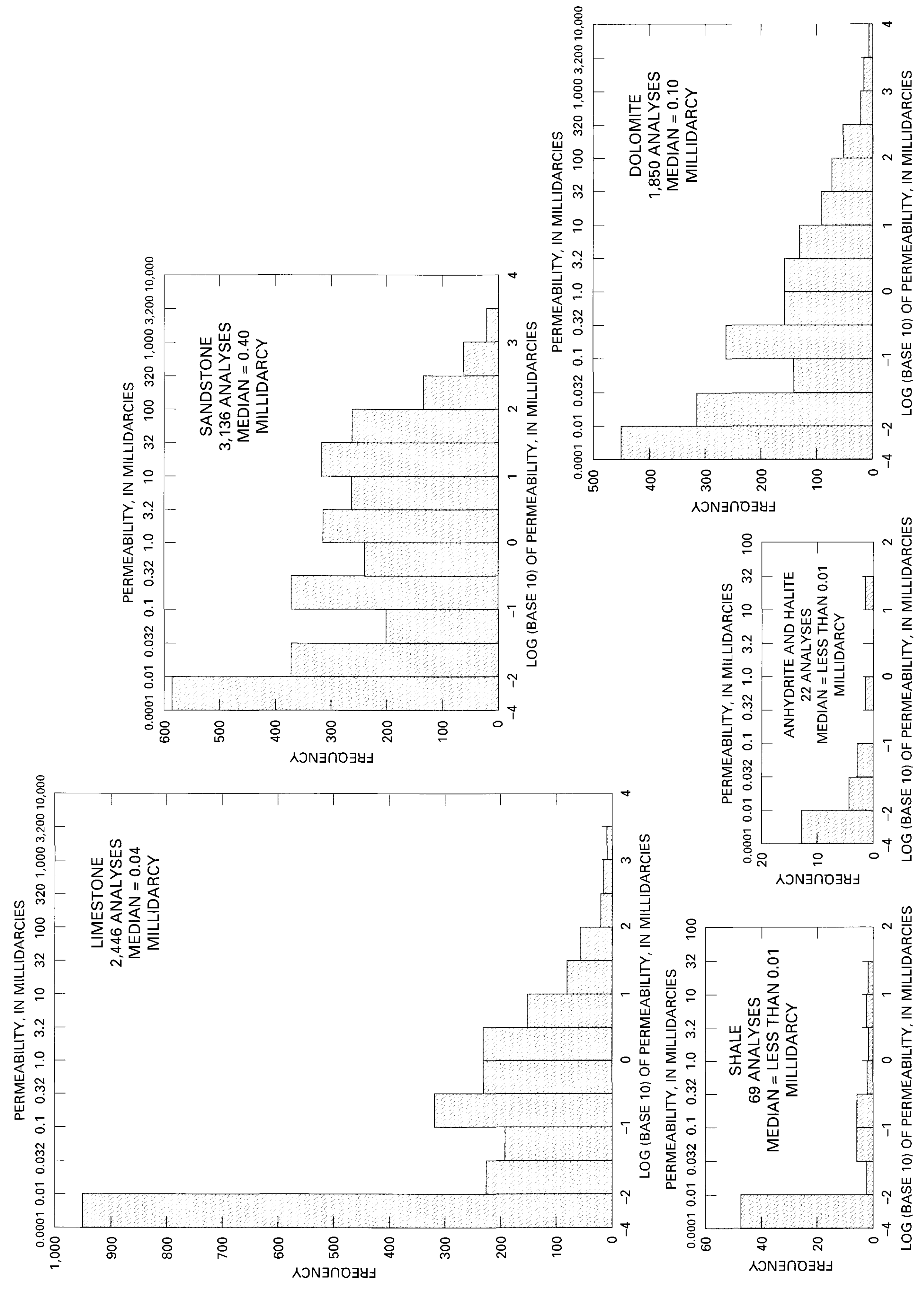


FIGURE 22.—Frequency distribution of pore-scale permeability in Paleozoic sedimentary rocks of the Upper Colorado River Basin.

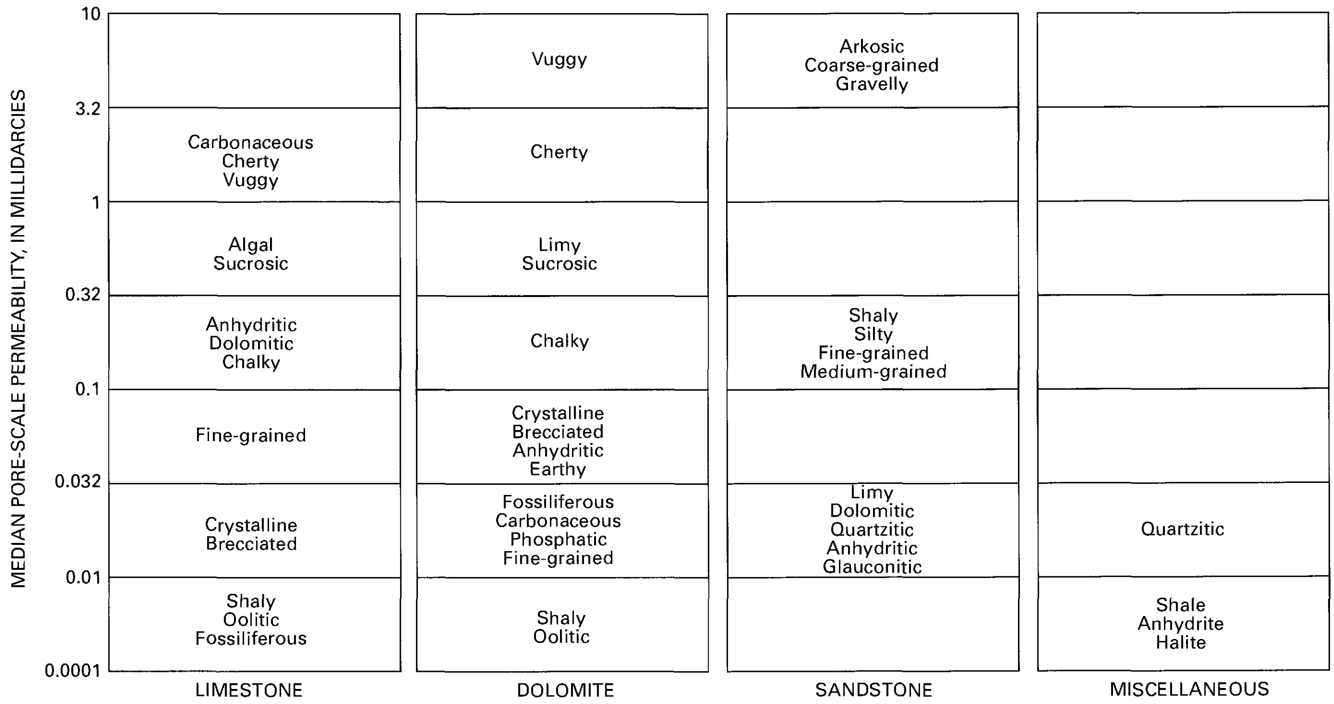


FIGURE 23.—Relation of pore-scale permeability to variations in the composition and texture of Paleozoic rocks of the Upper Colorado River Basin.

The Four Corners aquifer system (table 1) consists, in ascending order, of the Flathead aquifer, Gros Ventre confining unit, Bighorn aquifer, Elbert-Parting confining unit, and Madison aquifer. The Madison aquifer is subdivided into two lithologically distinct zones: the Redwall-Leadville zone and the Darwin-Humbug zone. Each of the aquifers, confining units, and zones, except the Madison aquifer, is named after component geologic units that typify the water-transmitting properties of the hydrogeologic unit. The name "Madison aquifer" is extended into the UCRB from the adjacent Northern Great Plains region (Downey, 1984). The Four Corners aquifer system is named for the importance of included aquifers in the four States that meet at the Four Corners—Arizona, Colorado, New Mexico, and Utah. The aquifer system name is not meant to be areally restrictive; included aquifers are equally important in the Wyoming part of the UCRB.

FLATHEAD AQUIFER

The Flathead aquifer consists of the Tintic Quartzite, Tapeats Sandstone and equivalents, Flathead Sandstone, lower part of the Lodore Formation, Sawatch Quartzite, and Ignacio Quartzite (table 1). Component geologic units are Early to Late Cambrian in age and become younger from west to east. The ability of these component geologic units to supply water to wells and springs depends mostly on the amount of cementation by silica and carbonate minerals and the degree of fracturing, which are related to past and present structural settings. In many areas, the component geologic units are so tightly cemented that they are not a dependable source of water. The Ignacio Quartzite in southwestern Colorado, for example, was considered by Whitfield and others (1983, table 3) to yield water only through fractures. In contrast, where component geologic units are friable to moderately cemented, they are reliable sources of water. The Flathead Sandstone in the Rawlins Uplift, for example, readily yields water to wells (Berry, 1960, p. 14). Because of these regional variations in water-supply capability, the Flathead aquifer can be expected to function as an aquifer subregionally in the UCRB.

THICKNESS AND LITHOLOGY

The Flathead aquifer is 0 to more than 800 ft thick (fig. 25). Throughout the region, the aquifer consists almost entirely of friable to quartzitic sandstone and quartzite, some of which is conglomeratic. In most areas, a basal conglomerate, a few feet to more than 50 ft thick, is present. Carbonate rocks account for 5 to 20 percent of the aquifer and increase toward depositional centers, particularly in northwestern Colorado, where the sandstone and quartzite commonly are dolomitic. Shale layers become more prevalent toward the upper part of the aquifer and can be nearly as abundant as sandstone and quartzite layers on the southern and eastern edges of the UCRB. Component

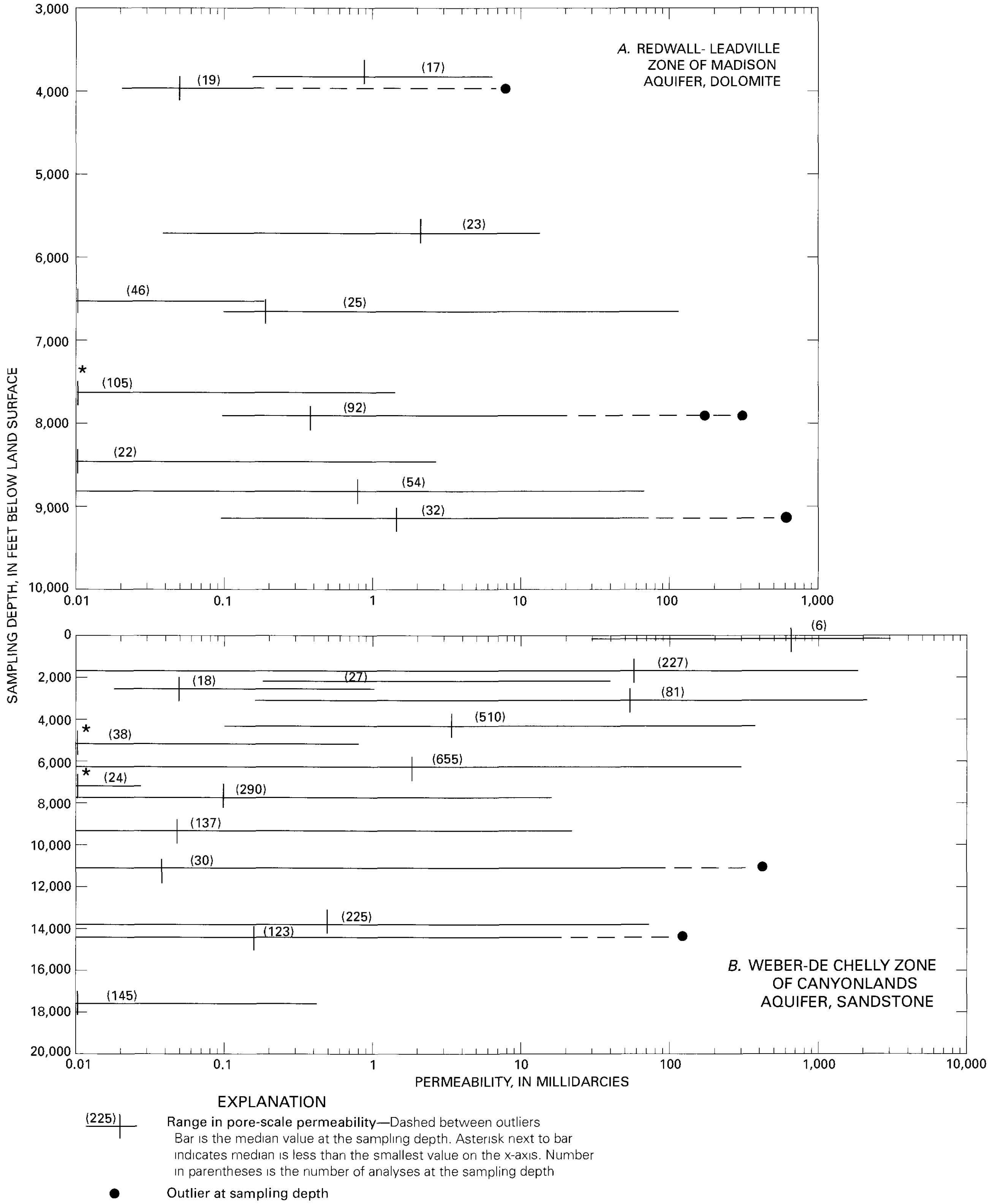


FIGURE 24.—Relation of pore-scale permeability to depth below land surface in the Redwall-Leadville zone of the Madison aquifer and Weber-De Chelly zone of the Canyonlands aquifer.

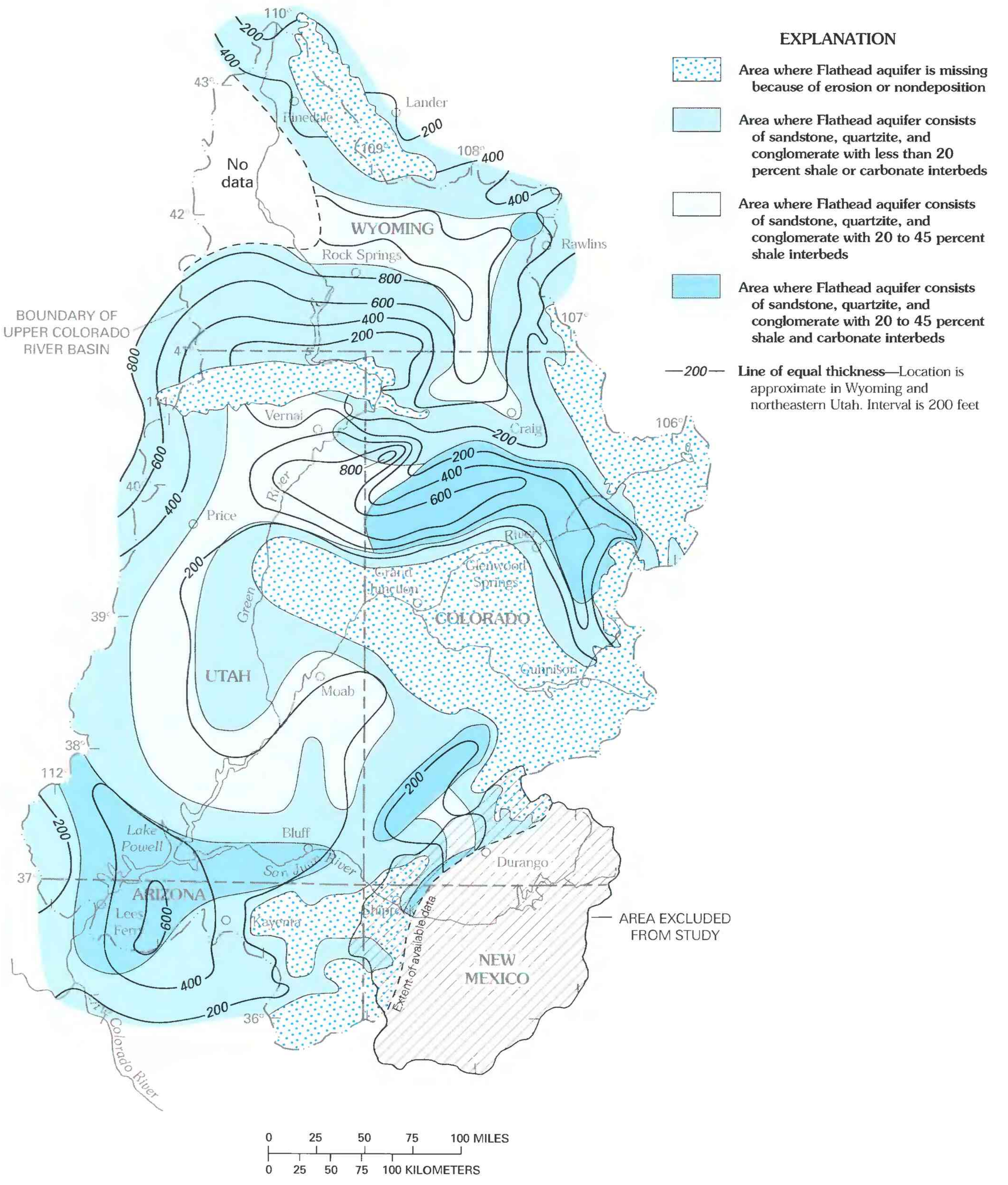


FIGURE 25.—Thickness and lithology of the Flathead aquifer. (Modified from Geldon, in press, pl. 6.)

geologic units in the Flathead aquifer are overlain in most areas conformably to gradationally by Cambrian shale and carbonate formations. However, in some areas, such as the eastern Great Divide Basin, eastern Uinta Mountains, and the Sawatch Range, component geologic units are overlain unconformably by carbonate and clastic rocks of Cambrian to Mississippian age.

POROSITY AND PERMEABILITY

Virtually no quantitative information regarding the porosity of the Flathead aquifer was found. Geophysically determined porosity in a borehole near Eagle, Colo., ranged from 0.5 to 24 percent, with a median value of 12 percent (fig. 26). Porosity values of 10 percent or less at depths of 10,945, 11,005, and 11,040 ft in this borehole probably correspond to intervals of dolomite that crop out in Glenwood Canyon nearby. Other variations in porosity may indicate lithologic changes from sandstone to quartzite. Because vertical variations in hydrologic properties tend to approximate regional variations, one can anticipate that unit-averaged porosity could vary regionally from less than 5 to more than 20 percent.

Measured values of local-scale permeability in this hydrogeologic unit vary by five orders of magnitude. Values of local-scale permeability determined from five drill-stem tests in the Great Divide, Piceance, and Paradox Basins ranged from 0.003 to 6.5 md, with a median value of 0.44 md. These values are typical of cemented sandstone with small to moderate permeability (fig. 19). In contrast, a drill-stem test on the Monument Upwarp indicated a local-scale permeability value of 42 md, typical of friable sandstone with moderate permeability. Regional trends cannot be extrapolated from the limited data.

HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

On the basis of the available data, regional variations in hydraulic conductivity appear to be substantial. In the Great Divide, Piceance, and Paradox Basins, hydraulic-conductivity values determined from five drill-stem tests ranged from 0.000007 to 0.016 ft/d, with a median value of 0.001 ft/d. In contrast to these small to moderate values, the hydraulic conductivity determined from a drill-stem test in an uplifted area, the Monument Upwarp, was 0.11 ft/d. The maximum hydraulic conductivity in uplifted areas is unknown, but it probably is at least as large as the hydraulic conductivity of similar strata in the underlying Uinta Mountain Group of Proterozoic age. Three pressure-injection tests of quartzite in the Uinta Mountain Group on the south flank of the Uinta Mountains by the Bureau of Reclamation (written commun., 1983) indicated hydraulic-conductivity values ranging from 0.29 to 0.40 ft/d. From an aquifer test nearby, Hood (1976, p. 63) estimated a hydraulic-conductivity value of 3 ft/d for the Uinta Mountain Group. If the values for the Proterozoic rocks are analogous, maximum values of hydraulic conductivity in the Flathead aquifer would appear to be between 0.1 and 3 ft/d, in the range of friable sandstone with large hydraulic conductivity.

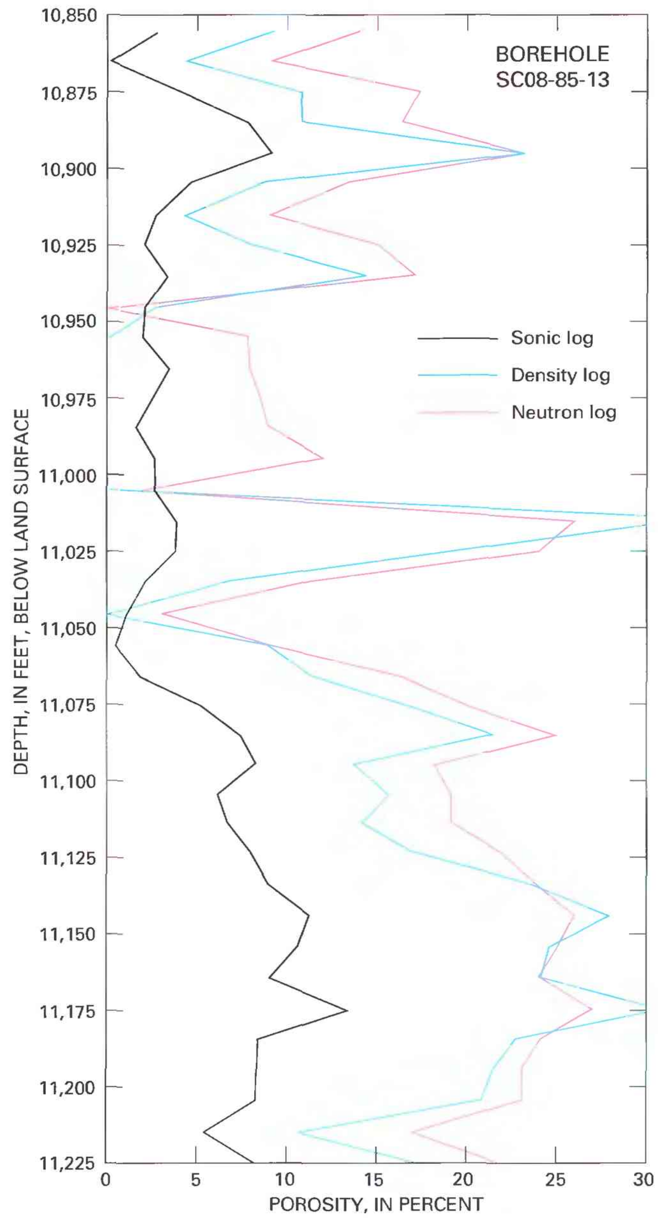


FIGURE 26.—Relation of geophysically determined porosity to depth below land surface in the Sawatch Quartzite near Eagle, Colorado.

The composite transmissivity of the Flathead aquifer probably ranges from 0.001 to 300 ft^2/d. Hydraulic-conductivity values determined from drill-stem tests and the thickness of the aquifer at the sites of these tests indicate a composite transmissivity range of 0.0011 to 41 ft^2/d (fig. 27). The previously mentioned aquifer test of the Uinta Mountain Group indicated a transmissivity of 190 ft^2/d. This latter value compares favorably with results of three flowing-well tests of the Flathead Sandstone on the faulted margin between the Bighorn Basin and the Bighorn Mountains (about 140 mi northeast of Pinedale, Wyo.) that were

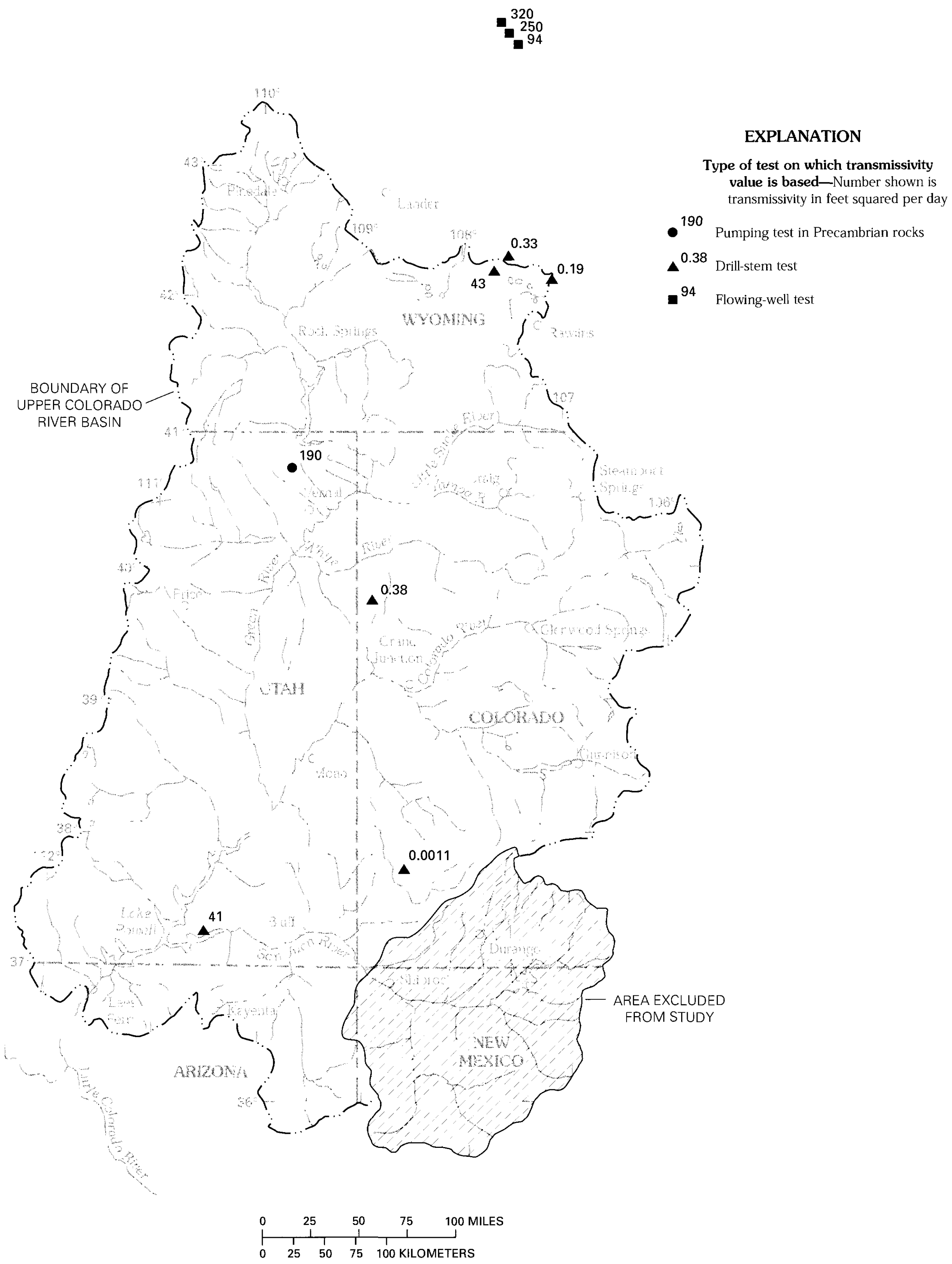


FIGURE 27.—Transmissivity distribution in the Flathead aquifer in the Upper Colorado River Basin and vicinity.

reported by Cooley (1985, p. 39). Transmissivity values determined from these tests ranged from 1 to 320 ft^2/d. Similarly large values of transmissivity might be possible where the Flathead aquifer is present on the flanks of fault-bounded uplifts in the UCRB.

YIELDS FROM WELLS AND SPRINGS

Variations in well and spring discharges reflect the wide recorded range in the transmissivity of the Flathead aquifer. In 10 drill-stem tests throughout the UCRB, yields ranged from 0.60 to 26 gal/min, with a median value of 3.8 gal/min. Similarly, springs issuing from the Tapeats Sandstone in the Grand Canyon were estimated by Metzger (1961, p. 128) to have discharges of less than 1 to 5 gal/min, and a spring issuing from the Sawatch Quartzite in the Elk Mountains (SC14–85–22dda) was determined by the U.S. Geological Survey (unpublished) to have a discharge of 25 gal/min. In contrast to these small discharges, a spring issuing from the Flathead Sandstone in the Rawlins Uplift discharges at a rate of 100 gal/min (Berry, 1960, p. 51). Two warm springs in the Elk Mountains that might be issuing from either the Sawatch Quartzite or overlying formations of Cambrian to Mississippian age also have moderate discharges. Cement Creek Warm Spring (SC14–84–18cac) discharges at rates of 60 to 80 gal/min from travertine deposits above the contact between Paleozoic rocks and Precambrian granitic rocks (Barrett and Pearl, 1977, p. 109) and may be related to a topographically and fault-controlled local flow system. Ranger Warm Spring (SC14–85–22dda) discharges at rates of 132 to possibly 250 gal/min (Barrett and Pearl, 1977, p. 112) and appears to be related to a fault that passes directly through the site of the spring. As indicated by flows from wells completed in the Flathead Sandstone in the Bighorn Basin (Cooley, 1985), discharges of 800 to 3,000 gal/min might be possible from the aquifer on the faulted margins between basins and uplifts within the UCRB.

GROS VENTRE CONFINING UNIT

The Gros Ventre confining unit consists of the Ophir Shale, Bright Angel Shale and equivalents, Gros Ventre Formation, and the upper part of the Lodore Formation (table 1). Component geologic units are Early to Late Cambrian in age and become younger from west to east. Geologic units included in this hydrogeologic unit generally are considered to have negligible permeability. For example, Lines and Glass (1975) state that the Gros Ventre Formation in the Overthrust Belt consists of "poorly permeable rock." Hood and Danielson (1981, p. 16) indicate that the Ophir Shale in the Henry Mountains Basin has "very low permeability." According to Metzger (1961, p. 116), "***the most important hydrologic characteristic of the Bright Angel Shale in the Grand Canyon area is the retardation of the downward percolation of ground water." Based on the consensus of opinion and limited data, this hydrogeologic unit was classified as a confining unit during this investigation.

THICKNESS AND LITHOLOGY

The Gros Ventre confining unit is 0 to 1,100 ft thick (fig. 28). Regionally, the confining unit is composed of claystone, siltstone, and sandy shale with subordinate interbeds of sandstone, limestone, and dolomite. The sandstone commonly is micaceous and glauconitic, but quartzitic layers are present near the base and margins of the confining unit. At the margins of depositional centers, sandstone and shale tend to be present in subequal amounts. Carbonate layers progressively increase in abundance toward depositional centers, where they can be as abundant as shale layers. From the Uinta Mountains south to the margins of the UCRB, component geologic units generally are overlain conformably to gradationally by Cambrian carbonate formations. North of the Uinta Mountains, however, the Gros Ventre Formation is overlain unconformably in most areas by carbonate and clastic rocks of Cambrian to Mississippian age.

PERMEABILITY AND HYDRAULIC CONDUCTIVITY

Because of its generally poor ability to supply water, the Gros Ventre confining unit virtually has been overlooked in hydrologic investigations. Consequently, little hydrologic data were obtained for this unit. No porosity information was found during this investigation. A drill-stem test of the Bright Angel Shale in the Paradox Basin (located at SLD 30–24–10da) indicated a local-scale permeability value of 0.18 md and a hydraulic-conductivity value of 0.00044 ft/d for an interval of interbedded sandstone, siltstone, and shale. In the western Uinta Mountains, pressure-injection tests of the Proterozoic Red Pine Shale, a unit similar to the Ophir Shale, may indicate upper limits of permeability and hydraulic conductivity for the Gros Ventre confining unit. A test of interbedded sandstone and siltstone in the Red Pine Shale indicated a local-scale permeability value of 10 md and a hydraulic-conductivity value of 0.025 ft/d; tests of fractured siltstone and shale indicated local-scale permeability values of 2.2 and 29 md and hydraulic-conductivity values of 0.006 and 0.07 ft/d (Bureau of Reclamation, written commun., 1983). Available data indicate that intervals within the Gros Ventre confining unit that either are fractured or contain substantial interbeds of sandstone probably have moderate permeability and hydraulic conductivity. Intervals that are unfractured or composed mostly of shale probably have small permeability and hydraulic conductivity. The limited data do not permit estimates of transmissivity.

YIELDS FROM WELLS AND SPRINGS

Although the Gros Ventre confining unit generally retards ground-water movement, discharges to wells and springs may occur from the more permeable layers. A drill-stem test in the

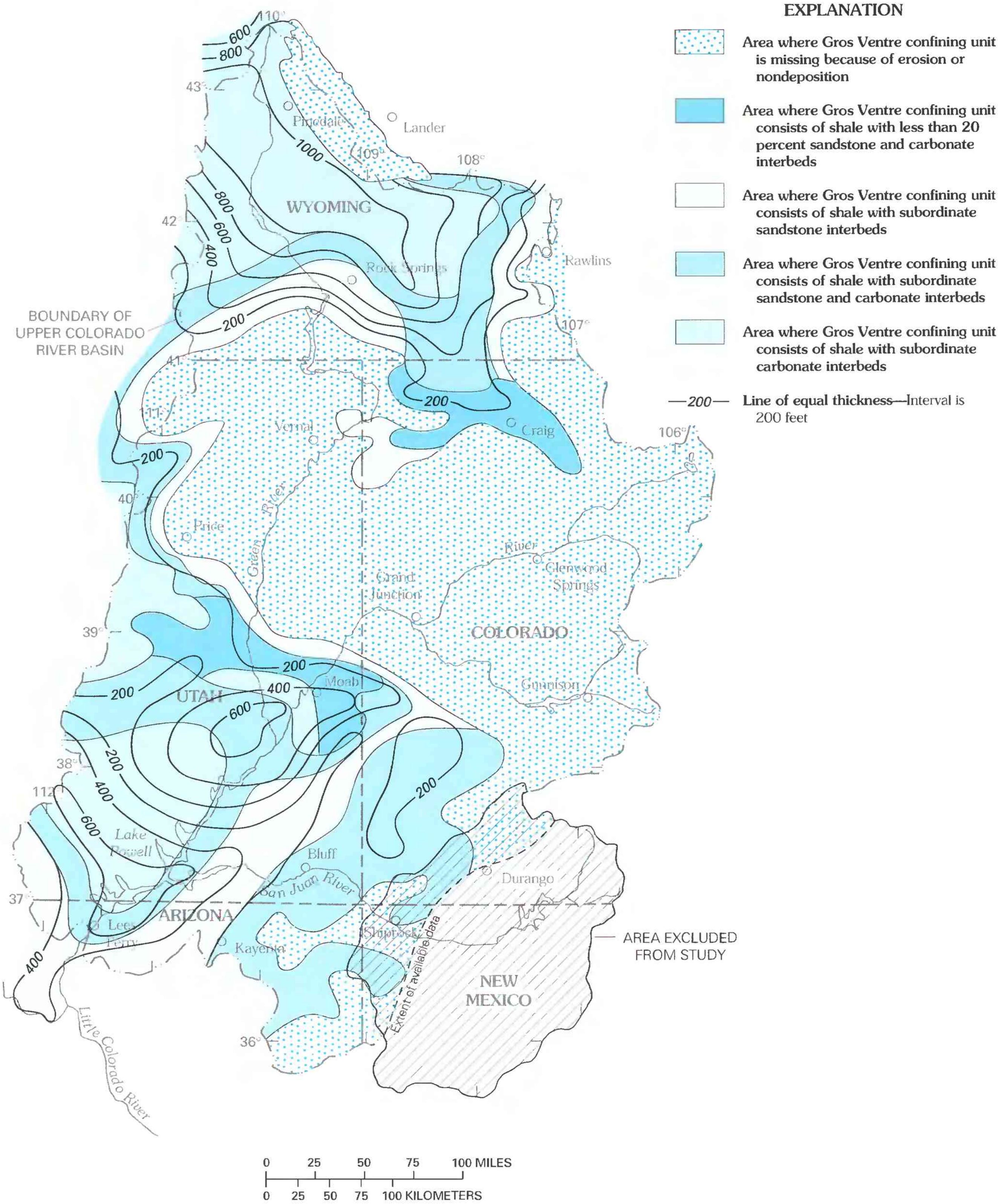


FIGURE 28.—Thickness and lithology of the Gros Ventre confining unit. (Modified from Geldon, in press, pl. 8.)

Paradox Basin induced a flow of 8 gal/min from the Bright Angel Shale. Four springs issuing from this formation in the Grand Canyon each have estimated discharges of less than 1 gal/min (Metzger, 1961, p. 128), and a spring issuing from the formation in the canyon of the Little Colorado River was estimated by McCulley (1985, p. 13) to have a discharge of 6 gal/min. These data indicate that typical yields from this confining unit probably do not exceed 25 gal/min. However, a spring (SB26–113–07db) near La Barge, Wyo., that issues from a sandy lens in the Park Shale Member of the Gros Ventre Formation discharges at an estimated rate of about 900 gal/min (Lines and Glass, 1975). According to Bertagnolli (1941, p. 1743), the water issuing from this spring probably originates in the Bighorn Dolomite and is forced to the surface where the stratigraphic and hydraulic continuity of the Bighorn is interrupted by a fault. The issuance of the spring from the Gros Ventre Formation, therefore, is coincidental and does not result from lateral movement of water through the confining unit.

BIGHORN AQUIFER

The Bighorn aquifer consists of the Muav Limestone and equivalents, Maxfield Limestone, Lynch Dolomite, Gallatin Limestone, Dotsero Formation, Peerless Formation, Manitou Dolomite, Harding Sandstone, Fremont Limestone, and Bighorn Dolomite (table 1). Component geologic units generally are considered to be Middle Cambrian to Late Ordovician in age. Water availability from this hydrogeologic unit appears to depend mostly on fractures and solution openings. In the Bighorn Dolomite in the Overthrust Belt, for example, there are large volumes of poorly permeable rock, but water is readily discharged from solution-enlarged fractures (Lines and Glass, 1975). In southeastern Utah, the Lynch Dolomite transmits water mostly through fractures and karst zones (Rush and others, 1982; Weir and others, 1983a). On the south rim of the Grand Canyon, the Muav Limestone is an aquifer because of the characteristic development of solution channels (Metzger, 1961, p. 117). The fractures and solution channels that make this hydrogeologic unit an aquifer in and near uplifted areas probably are not well developed in structural basins, and, in these latter areas, the unit may be too intact to transmit much water. Shale layers, which are abundant locally, also would impede ground-water movement. Because of these regional variations in water-supply capability, this hydrogeologic unit can be expected to function as an aquifer subregionally in the UCRB.

THICKNESS AND LITHOLOGY

The Bighorn aquifer is 0 to more than 3,000 ft thick (fig. 29). Regionally, the aquifer consists of limestone and dolomite with subordinate shale layers and generally less than 5 percent sandstone layers. The Cambrian carbonate rocks within the Bighorn aquifer typically are glauconitic and oolitic and contain flat-pebble or edgewise conglomerate layers; in the Four Corners area, the rocks are shaly. The Ordovician carbonate rocks within the aquifer can be granular, crystalline, or fine grained, and the Manitou Dolomite tends to be siliceous. Shale and sandstone layers account for less than 20 percent of the aquifer and are most abundant toward its bottom and margins, where component geologic units grade into other Cambrian formations. In all areas, the geologic units that compose the Bighorn aquifer are overlain unconformably by Devonian and Mississippian rocks.

POROSITY AND PERMEABILITY

On the basis of samples mostly from the Muav Limestone in the Four Corners area, porosity in the Bighorn aquifer is estimated to range from 0.1 to 13 percent. In 46 samples of dolomite from three borehole intervals in the Muav Limestone, the porosity ranged from 0.1 to about 13 percent, with a median value of 1.8 percent (table 6). In a borehole near Big Piney, Wyo., the Bighorn Dolomite exhibited a range in geophysically determined porosity of 0.6 to 9.3 percent and a median porosity of 5.0 percent (fig. 30). The unit-averaged porosity at the four data sites, where data is obtained, is estimated to range from 1.5 to 5.0 percent.

The available data indicate that the Bighorn aquifer characteristically exhibits moderate permeability. In 56 samples of dolomite and sandstone from the Muav Limestone, pore-scale permeability ranged from less than 0.01 to 157 md, with median values of less than 0.01 md for the dolomite and 0.85 md for the sandstone (table 6). For the dolomite, there appears to be a crude relation between porosity and pore-scale permeability (fig. 31). Using this relation (for which an equation is given in table 3), the Bighorn Dolomite near Big Piney, Wyo., was estimated to have a unit-averaged pore-scale permeability value of 0.13 md, which, according to equation 8, is equivalent to a local-scale permeability value of 0.68 md. In comparison, 10 drill-stem tests of beds equivalent to the Muav Limestone and Lynch Dolomite in southeastern Utah and northeastern Arizona indicated values of local-scale permeability ranging from 0.029 to 23 md; the median of these values was 2.9 md. The available data are insufficient for extrapolation of regional trends in permeability.

HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

The permeability data discussed in the previous section are equivalent to a range in hydraulic conductivity of 0.000071 to 0.056 ft/d and a median hydraulic conductivity of 0.007 ft/d. As shown in figure 29, unit-averaged hydraulic conductivity in the Bighorn aquifer in the Four Corners area ranges from 0.00075 to 0.056 ft/d, and near Big Piney, Wyo., the unit-averaged hydraulic conductivity is 0.0017 ft/d.

On the basis of the available values of unit-averaged hydraulic conductivity and the thickness of the Bighorn aquifer at sites where these hydraulic-conductivity values were determined, the range in composite transmissivity for the aquifer is

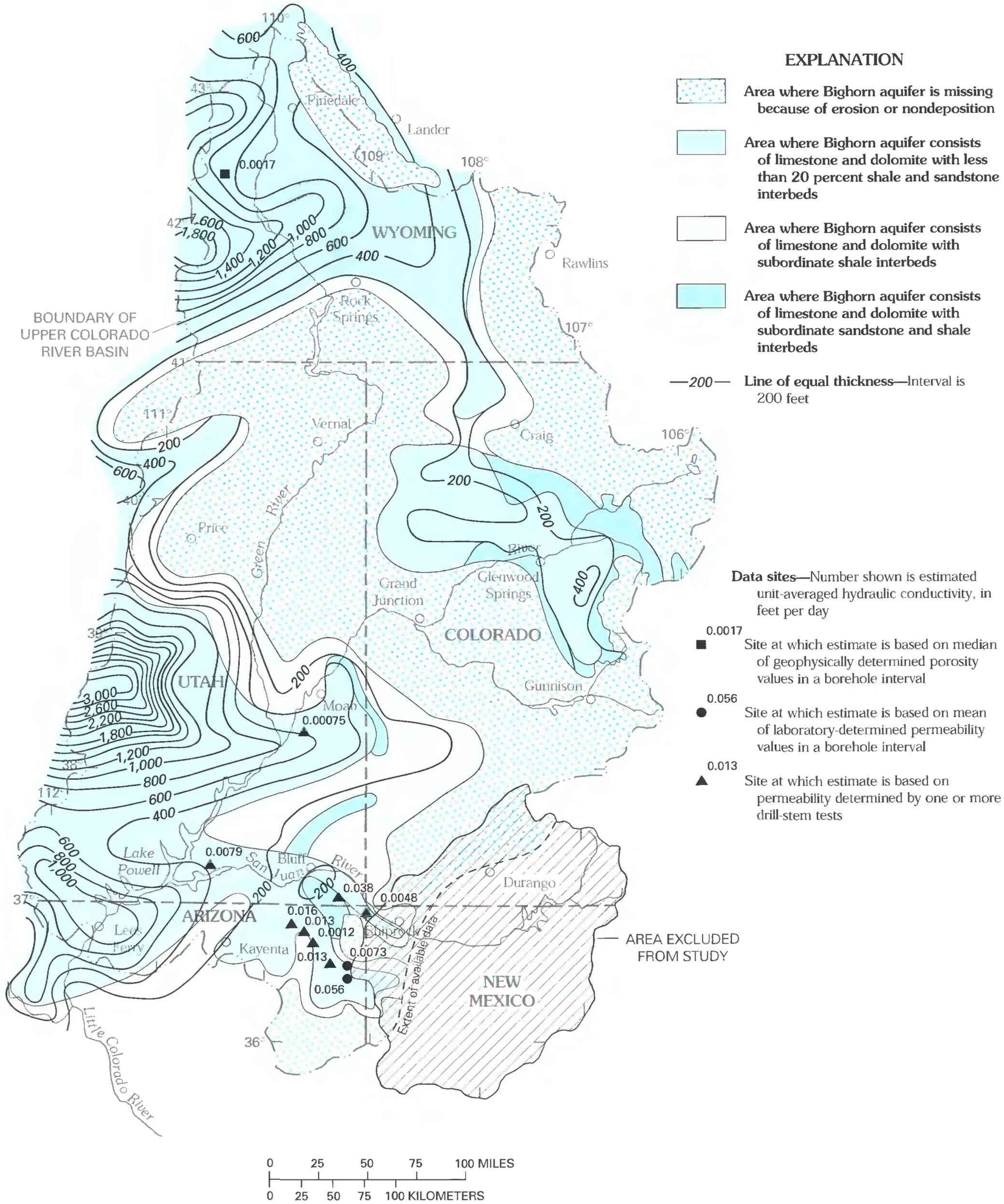


FIGURE 29.—Thickness, lithology, and unit-averaged hydraulic conductivity of the Bighorn aquifer. (Modified from Geldon, in press, pl. 9.)

TABLE 6.—*Porosity and pore-scale permeability statistics for the Bighorn aquifer*

[In this and all subsequent compilations of porosity and permeability statistics, the number of observations for a rock type does not equal the sum of observations for subcategories because (1) many samples of a rock type were not classifiable into a subcategory; and (2) many samples were classifiable into more than one subcategory. For example, a sample described as "shaly, fine-grained sandstone" could be included in the statistical analyses of sandstone, shaly sandstone, and fine-grained sandstone. Samples that were described as combinations of mutually exclusive rock types, such as "limestone and shale" were rejected for the statistical analysis of either rock type. Samples that were described as combinations of mutually exclusive textures, such as "crystalline to sucrose limestone," were included in the statistical analysis for the rock type but were rejected for the statistical analysis of either texture; <, less than]

Rock type	Porosity (percent)			Number of observations	Pore-scale permeability (millidarcies)			Number of observations
	Minimum	Maximum	Median		Minimum	Maximum	Median	
Dolomite								
Shaly	0.1	8.1	2.1	20	<0.01	157	<0.01	20
Vuggy	1.7	12.6	3.5	9	.01	157	1.6	9
Medium-grained	7.0	12.6	11	4	.10	157	3.6	4
All	0.1	12.6	1.8	46	<.01	157	<.01	46
Sandstone	2.5	8.2	4.6	10	.01	5.5	.85	10

estimated to be at least 0.10 to 6 ft^2/d. Composite transmissivity values probably are larger than 6 ft^2/d in uplifted areas where the Bighorn aquifer is faulted and yields large quantities of water to springs (see discussion in next section). Composite transmissivity values probably are smaller than minimum estimates where the Bighorn aquifer is sparsely fractured (in the interiors of structural basins) or contains abundant shale interbeds.

YIELDS FROM WELLS AND SPRINGS

Yields to most wells and springs from the Bighorn aquifer are less than 500 gal/min. In 13 drill-stem tests of the Muav Limestone, Lynch Dolomite, and Bighorn Dolomite, discharges ranged from 3.0 to 41 gal/min, with a median value of 7.5 gal/min. Numerous springs and seeps issuing from the Muav Limestone in the Grand Canyon have discharges ranging from less than 1 to 10 gal/min, and collectively, the springs at Hermit Creek and Indian Gardens discharge at rates of 200 to 300 gal/min (Metzger, 1961; Cooley, 1976). In the Overthrust Belt, three springs issuing from the Bighorn Dolomite have discharges of 125 to 450 gal/min. These data indicate that under most circumstances, yields from the Bighorn aquifer are likely to be in the small to moderate range. Near faults, however, large discharges may be possible. As previously mentioned, a spring near La Barge, Wyo., with a discharge of about 900 gal/min probably is caused by water traveling along a fault from the Bighorn Dolomite to a permeable bed in the Gros Ventre Formation. Another spring in the Overthrust Belt (SB34–118–26aad) discharges from the Bighorn Dolomite at a rate of 3,200 gal/min (Lines and Glass, 1975). The location of this spring, too, is controlled by a fault, and some of the flow probably is from the Lodgepole and Mission Canyon Limestones of the overlying Redwall-Leadville zone of the Madison aquifer.

ELBERT-PARTING CONFINING UNIT

The Elbert-Parting confining unit consists of the Darby, Parting, Elbert, and Temple Butte Formations, and the Cottonwood Canyon Member of the Lodgepole and Madison Limestones (table 1), which are Late Devonian to Early Mississippian in age. The confining unit is so heterogeneous that its capability to supply water mostly depends on the rock types that are present in an area and, secondarily, on whether carbonate rocks contain solution channels and(or) whether the rocks, in general, are fractured. Large parts of the Elbert Formation, for example, consist of negligibly permeable rock; but in the Four Corners area, sandstone and carbonate beds in the lower part of the formation produce water (INTERA Environmental Consultants, Inc., 1984). The Darby Formation does not transmit water readily to wells, but near faults in the Overthrust Belt, springs issue from the formation (Lines and Glass, 1975). Similarly, the Parting Formation generally is not known to transmit water, but in a well that was drilled at the fault-bounded southern end of the Park Range (SC02–84–03), the flow from the well apparently increased when fractured quartzite in the lower part of the Parting Formation was penetrated (unpublished consultant's report furnished by the Colorado Division of Water Resources, written commun., 1984). Regionally, the hydrogeologic unit impedes flow in more places than it transmits water and, therefore, it was classified as a confining unit in this investigation.

THICKNESS AND LITHOLOGY

The thickness of the Elbert-Parting confining unit ranges from 0 to more than 1,000 ft (fig. 32). Regionally, the confining unit is characterized by highly variable proportions of limestone, dolomite, sandstone, quartzite, shale, and anhydrite. Generally, the Darby, Elbert, and Temple Butte Formations gain carbonate rocks at the expense of clastic layers westward across the UCRB, and the Parting Formation becomes more carbonaceous toward the center of its depositional area. In most areas, carbonate rocks comprise at least 50 percent of the confining unit. Shale layers generally account for at least one-third of the confining unit and locally exceed carbonate layers in abundance. Sandstone and quartzite layers rarely compose more than one-third of the confining unit, but they predominate on the southeastern edge of the Darby–Cottonwood Canyon depositional area, on the northeastern edge of the Parting depositional area, and at the base of the Elbert Formation. Anhydrite layers account for as much as 10 percent of the confining unit in the Overthrust Belt west of Kemmerer, Wyo. Component geologic units in

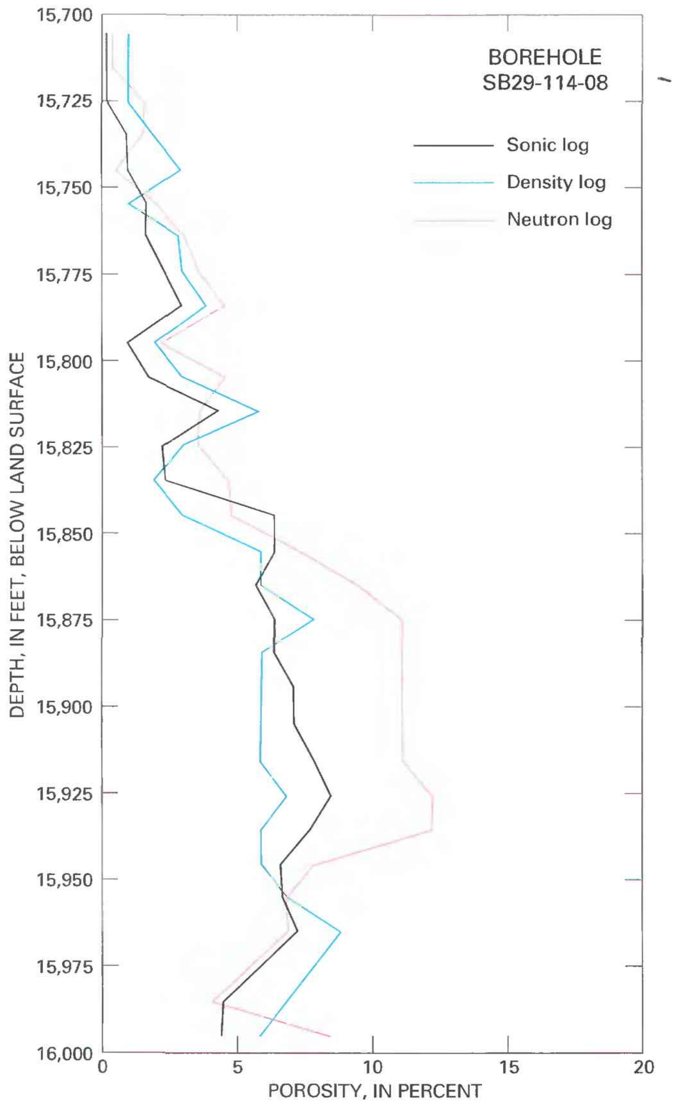


FIGURE 30.—Relation of geophysically determined porosity to depth below land surface in the Bighorn Dolomite near Big Piney, Wyoming.

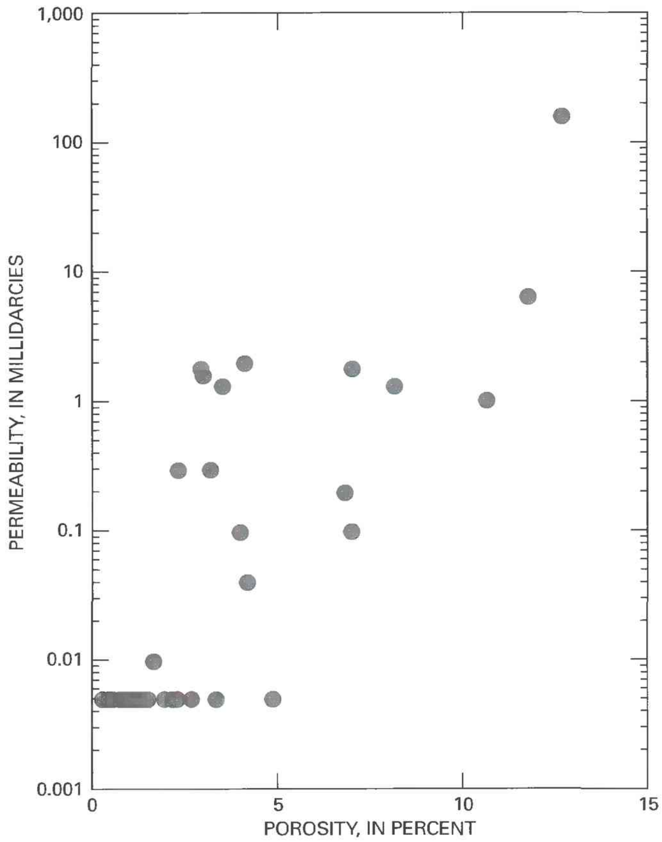


FIGURE 31.—Relation of porosity to pore-scale permeability in dolomite samples from the Muav Limestone.

Wyoming and the Grand Canyon area are overlain unconformably by Mississippian carbonates, whereas the Elbert and Parting Formations generally are overlain conformably by formations of Devonian to Mississippian age.

POROSITY AND PERMEABILITY

The only porosity data found for the Elbert-Parting confining unit were for the Elbert Formation, but on the basis of the lithologic similarity of the Elbert Formation to the other component geologic units, it is presumed that these data are representative of the entire confining unit. The available data indicate a porosity range of 0.3 to about 12 percent for rock types within the confining unit (fig. 33 and table 7). Sandstone, with a median porosity of 6.1 percent, is more porous than dolomite, with a median porosity of 1.9 percent (table 7). On the basis of relatively few analyses, shale and quartzite both appear to have median porosity values of about 1 percent. On the basis of lithologic composition and median porosity values for the common rock types that are present, unit-averaged porosity for the Elbert-Parting confining unit is estimated to range regionally from less than 2 to about 5 percent (fig. 34).

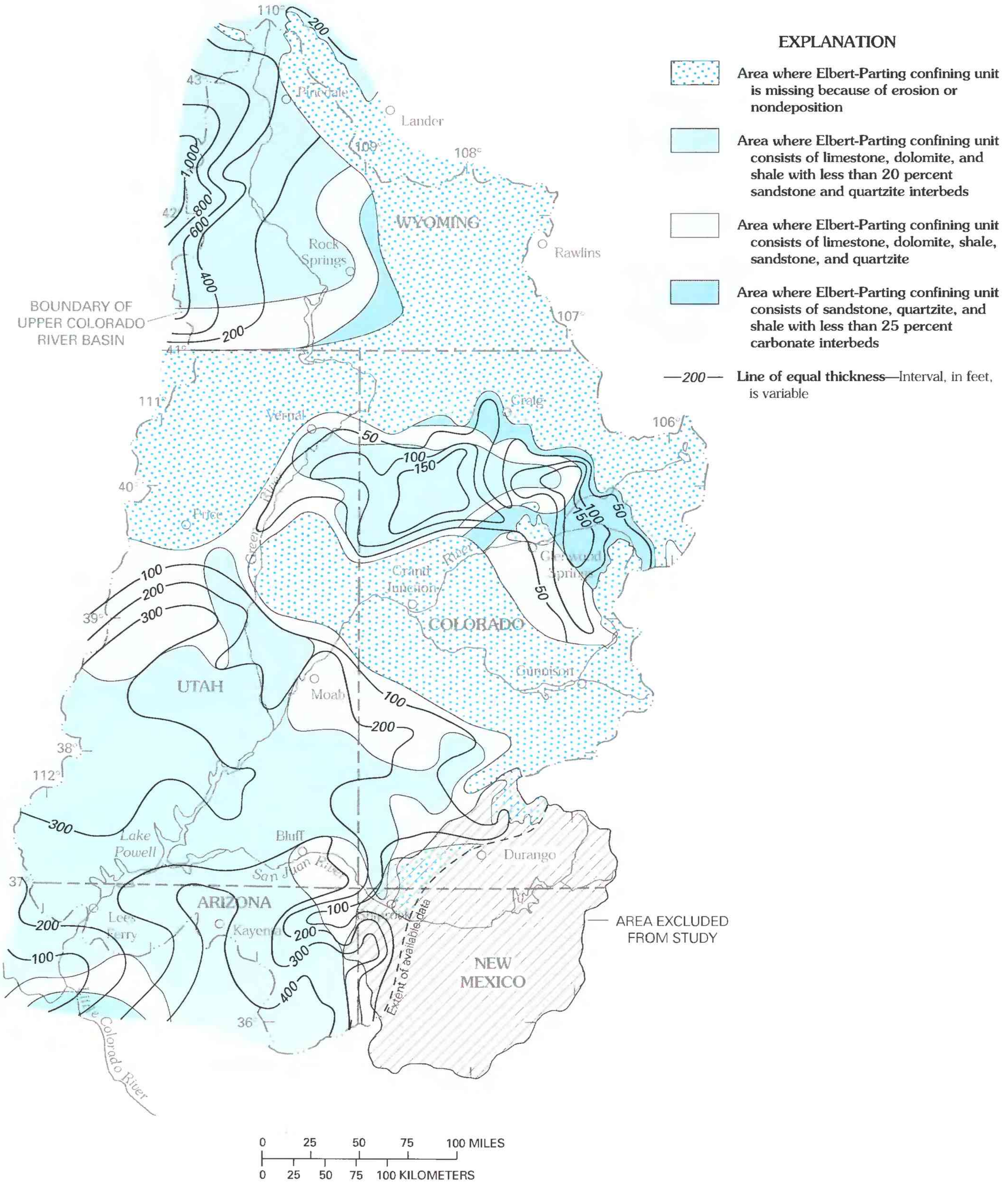


FIGURE 32.—Thickness and lithology of the Elbert-Parting confining unit. (Modified from Geldon, in press, pl. 10.)

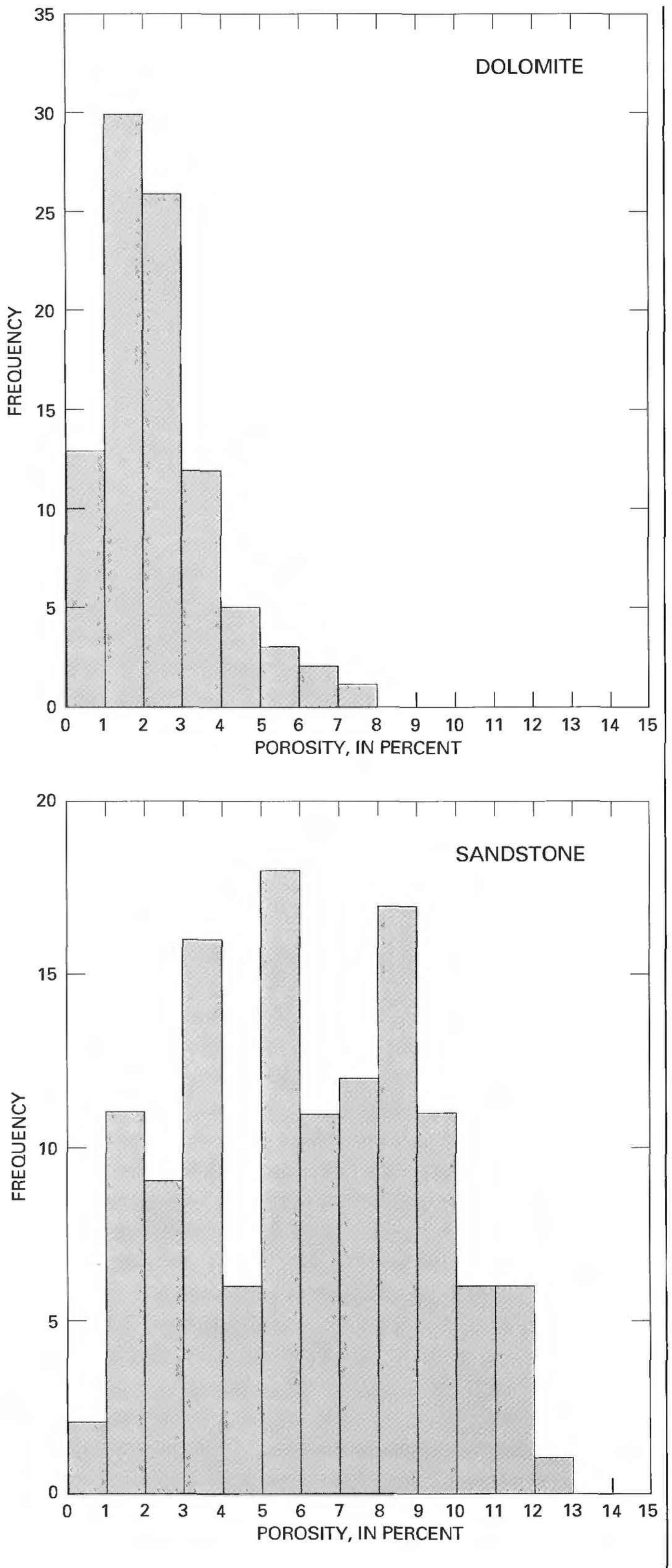


FIGURE 33.—Frequency distribution of porosity in samples of dolomite and sandstone from the Elbert Formation.

Interpretations of permeability in the Elbert-Parting confining unit also are based entirely on information for the Elbert Formation. These data indicate that pore-scale permeability ranged from less than 0.01 to 179 md, with median values of 0.42 md for sandstone, 0.02 md for quartzite, and less than 0.01 md for dolomite and shale (table 7 and fig. 35). Pore-scale permeability is crudely related to porosity in sandstone samples (fig. 36) but is not related to porosity in dolomite samples (table 3). In 30 drill-stem tests, local-scale permeability ranged from 0.021 to 95 md, with median values of 2.3 md in intervals without shale and 1.2 md in intervals with shale. Together, the laboratory and field data indicate that the confining unit has small to moderate permeability, which is highly dependent on proportions of sandstone, dolomite, and shale.

HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

The permeability data discussed in the previous section are equivalent to a range in hydraulic conductivity of 0.000051 to 0.25 ft/d (fig. 37). The median of these values is 0.004 ft/d and is twice as large for intervals without shale (0.006 ft/d) than for intervals with shale (0.003 ft/d). Unit-averaged hydraulic conductivity of the Elbert Formation, based on tests of lithologically representative intervals, is estimated in the Four Corners area to range from 0.00005 to 0.25 ft/d, increasing from basins to uplifted areas (fig. 38).

The composite transmissivity of the Elbert Formation, based on areal distributions of unit-averaged hydraulic conductivity and thickness of the formation, is estimated to range from 0.008 to 200 ft^2/d (fig. 39). The transmissivity of this formation probably is as large as that of any other component geologic unit in the Elbert-Parting confining unit because the Elbert Formation is the most consistently water bearing of these geologic units. Therefore, although the minimum values of composite transmissivity for the Elbert-Parting confining unit may be less than that for the Elbert Formation, it is not likely that the confining unit is more transmissive where it is composed of geologic units other than the Elbert Formation, except possibly along thrust faults in the Overthrust Belt. Tentatively, the range in composite transmissivity for this confining unit is estimated to be that of the Elbert Formation.

YIELDS FROM WELLS AND SPRINGS

Yields from the Elbert-Parting confining unit typically are negligible to moderate. In the Four Corners area, many drill-stem tests produced negligible quantities of water, but 29 tests produced discharges ranging from 1.5 to 58 gal/min. The median value for these tests was 8.5 gal/min. In the Overthrust Belt, one of three springs known to be issuing from the Darby Formation (SB33–117–24b) has a discharge of 40 gal/min (Lines and Glass, 1975), which is consistent with discharges from the Elbert Formation. However, two springs in the

TABLE 7.—*Porosity and pore-scale permeability statistics for the Elbert-Parting confining unit*

[<, less than]

Rock type	Porosity (percent)			Number of observations	Pore-scale permeability (millidarcies)			Number of observations
	Minimum	Maximum	Median		Minimum	Maximum	Median	
Dolomite								
Crystalline	0.5	7.4	1.7	56	<0.01	7.0	<0.01	56
Shaly	.4	5.2	2.0	54	<.01	.36	<.01	54
Sandy	.4	3.7	2.2	35	<.01	<.01	<.01	35
All	.4	7.4	1.9	92	<.01	7.0	<.01	92
Sandstone					<.01			
Glauconitic	2.0	6.5	5.0	7	<.01	.04	.01	7
Shaly	.8	8.5	5.3	23	<.01	62	.10	23
Quartzitic or dolomitic	1.1	11.9	6.9	36	<.01	24	1.0	36
Fine-grained	.8	12.2	6.5	83	<.01	179	.81	83
Medium-grained	1.1	10.6	5.7	18	<.01	7.0	.32	18
Coarse-grained	2.7	8.4	5.3	12	<.01	55	1.6	12
All	.8	12.2	6.1	121	<.01	179	.42	121
Quartzite	1.0	4.1	1.2	5	<.01	.11	.02	5
Shale	.3	2.5	1.4	5	<.01	.05	<.01	5

Overthrust Belt (SB38–115–03bca and SB38–115–03bcd) have discharges of 900 and 1,100 gal/min (Lines and Glass, 1975). These two large springs issue in a zone of thrust faults and could be outlets for water traveling along fault planes from the Lodgepole and Mission Canyon Limestones. As indicated in figure 40, the two large springs clearly are outliers and, consequently, they are not considered representative of the water-bearing properties of the Elbert-Parting confining unit.

REDWALL-LEADVILLE ZONE OF THE MADISON AQUIFER

The Redwall-Leadville zone of the Madison aquifer consists of the Dyer Dolomite, Ouray Limestone, Gilman Sandstone, Leadville Limestone, and Redwall Limestone; the main bodies of the Lodgepole and Madison Limestones; and the lower part of the Mission Canyon Limestone (table 1). Component geologic units mostly are Late Devonian to Late Mississippian in age. Although pore-scale permeability in the Redwall-Leadville zone is relatively small, the zone transmits water through a system of interconnected solution channels and fractures nearly everywhere that it occurs. Examples include the Madison Limestone in the Rawlins Uplift (Berry, 1960) and Uinta Mountains (Hood, 1976); the Leadville Limestone in the White River Plateau (Teller and Welder, 1983), and the Redwall Limestone in the San Rafael Swell (Hood and Patterson, 1984) and Grand Canyon (Metzger, 1961). The relatively large and omnipresent permeability of the Redwall-Leadville zone, according to Lines and Glass (1975), exists because some of the solution channels in the aquifer developed during Pennsylvanian uplift of the region. Older carbonate units do not possess these ancient solution features and have solution permeability only near modern uplifted areas. In contrast, the Redwall-Leadville zone, because of its relict solution channels, can be permeable even where it is deeply buried. Indeed, the Redwall-Leadville zone in the Paradox Basin is more permeable than all other Paleozoic and Mesozoic rocks except some Jurassic sandstone formations (Rush and others, 1982; Weir and others, 1983a; Whitfield and others, 1983).

The importance of fracturing and solution channels to flow through the Redwall-Leadville zone is demonstrated by interception of streamflow at outcrops of the Madison Limestone. In Sinks Canyon on the north flank of the Wind River Mountains, the entire flow of the Middle Popo Agie River disappears into a cavern in the Madison Limestone, emerging about half a mile downstream as springs from the Tensleep Sandstone (Whitcomb and Lowry, 1968). On the south flank of the Uinta Mountains, the entire flows of Big Brush Creek, Pole Creek, and the three branches of the Dry Fork of Ashley Creek disappear into caves or sinkholes in the Madison Limestone. Springs issuing from the Madison Limestone, Pennsylvanian and Permian Weber Sandstone, Permian Park City Formation, and Triassic Moenkopi Formation downstream from these sinks have been traced by dye tests to the sinks (Maxwell and others, 1971). Springs issuing in Big Brush Creek and Pole Creek are traceable to sinks directly upstream. However, water disappearing into sinks on the Dry Fork of Ashley Creek emerges as springs not only in the Dry Fork but also in Deep Creek and the main fork of Ashley Creek (see figs. 1 and 2 of Maxwell and others, 1971). The tracer tests confirmed long-held

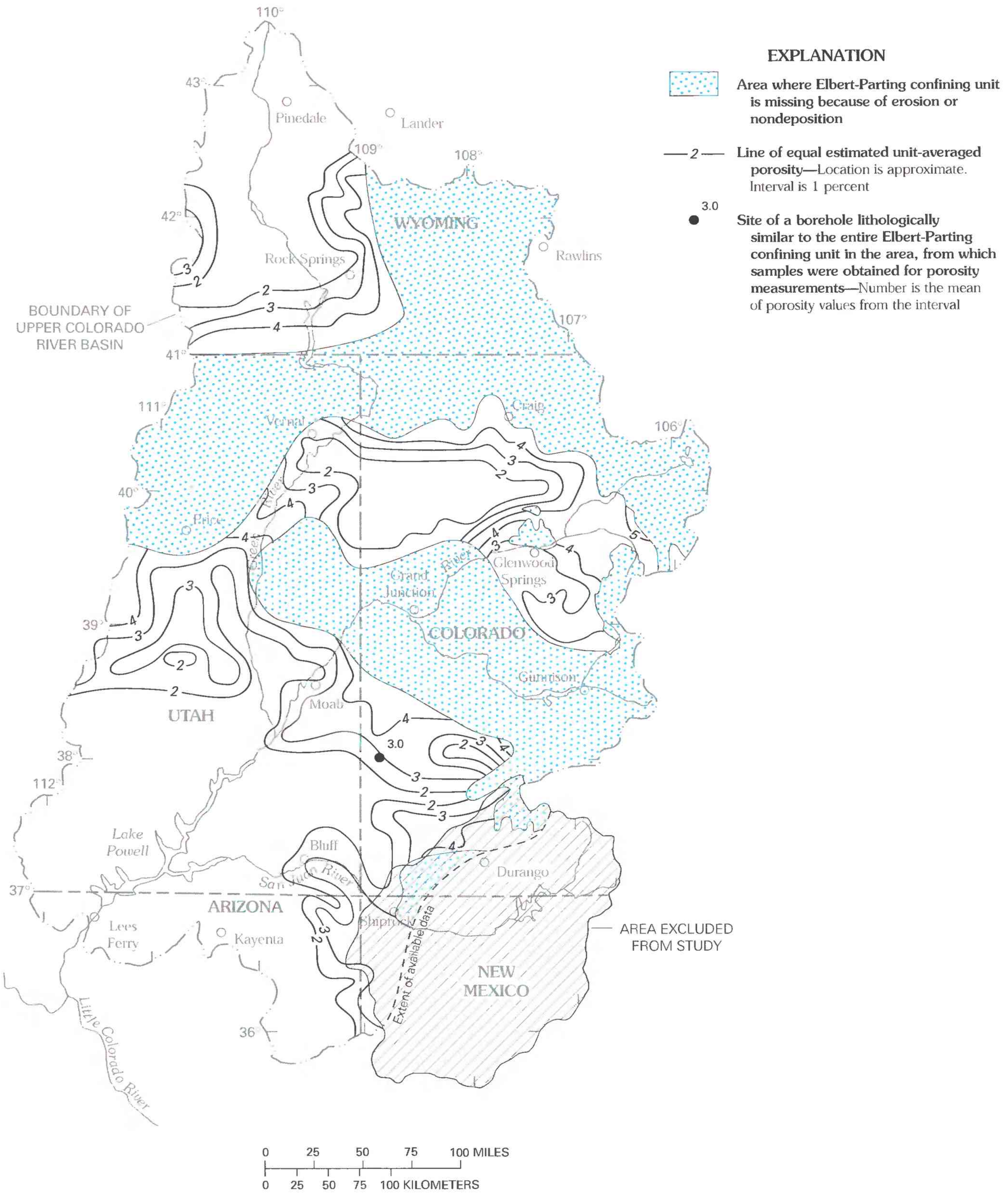


FIGURE 34.—Estimated distribution of unit-averaged porosity in the Elbert-Parting confining unit.

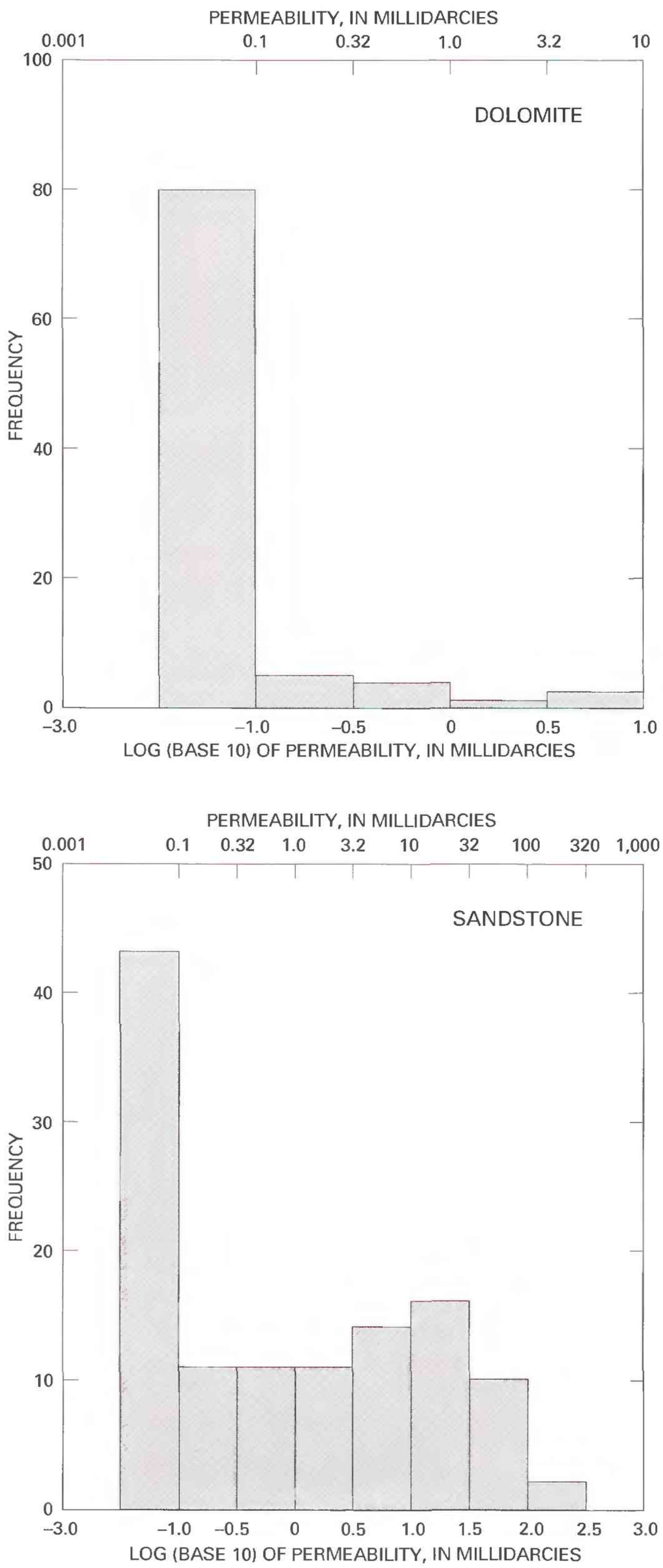


FIGURE 35.—Frequency distribution of pore-scale permeability in samples of dolomite and sandstone from the Elbert Formation.

hypotheses that a well-developed system of interconnected fractures and solution channels exists in the Madison Limestone and, by extension, other limestone formations comprising the Redwall-Leadville zone. Therefore, the Redwall-Leadville zone can be expected to function as an aquifer throughout the UCRB.

THICKNESS AND LITHOLOGY

The thickness of the Redwall-Leadville zone ranges from 0 to 2,500 ft (fig. 41). Limestone, dolomite, and bedded and nodular chert compose more than 95 percent of the Redwall-Leadville zone, with the proportion of limestone to dolomite decreasing away from uplifted areas. Regionally, careful tracing of stratigraphic markers between outcrops and boreholes indicates that limestone beds grade laterally into dolomite beds, such that neither the proportions nor sequence of limestone and dolomite can be used to establish contacts between Devonian and Mississippian formations. Sandstone and shale, concentrated in the interval between Devonian and Mississippian carbonate rocks and at the top of the Mississippian sequence, account for 5 to 10 percent of the Redwall-Leadville zone in areas interpreted to have been sites of shallow marine or intermittently subaerial deposition. The upper surface of the Redwall-Leadville zone is an unconformity that progressively truncates older rocks from west to east across the region. Shale layers, channel sandstone, or a rubble of carbonate and chert blocks in a matrix of shale or sandstone generally is present above the unconformity.

POROSITY AND PERMEABILITY

The porosity of carbonate rocks in the Redwall-Leadville zone depends on both composition and texture. Limestone porosity ranged from 0.3 to 15 percent, with a median value of 1.7 percent (fig. 42). Dolomite, on the average, is nearly four times more porous than limestone. Dolomite porosity ranged from 0.3 to about 22 percent, with a median value of 6.5 percent (fig. 42). In all types of limestone, median porosity ranged from about 1 to 3 percent (table 8). Dolomite porosity clearly is larger in varieties with a tendency for solution cavities to develop around inclusions, such as cherty, anhydritic, or vuggy dolomite, than in laminar varieties, such as shaly dolomite (table 8). On the other hand, a study by Thayer (1983) of the Madison Limestone in and near the Powder River Basin of Montana and Wyoming found that porosity and pore-scale permeability in this formation can be reduced substantially by secondary anhydrite, chert, and calcite in pores and fractures. Thayer (1983, fig. 24) also found that porosity in the Madison Limestone is independent of sampling depth, except possibly at depths of 3,000 ft or less. Because of lithologic variation and, possibly, variations in fracture development, porosity in the Redwall-Leadville zone of the Madison aquifer in the UCRB also shows no systematic variation with depth (for example, see

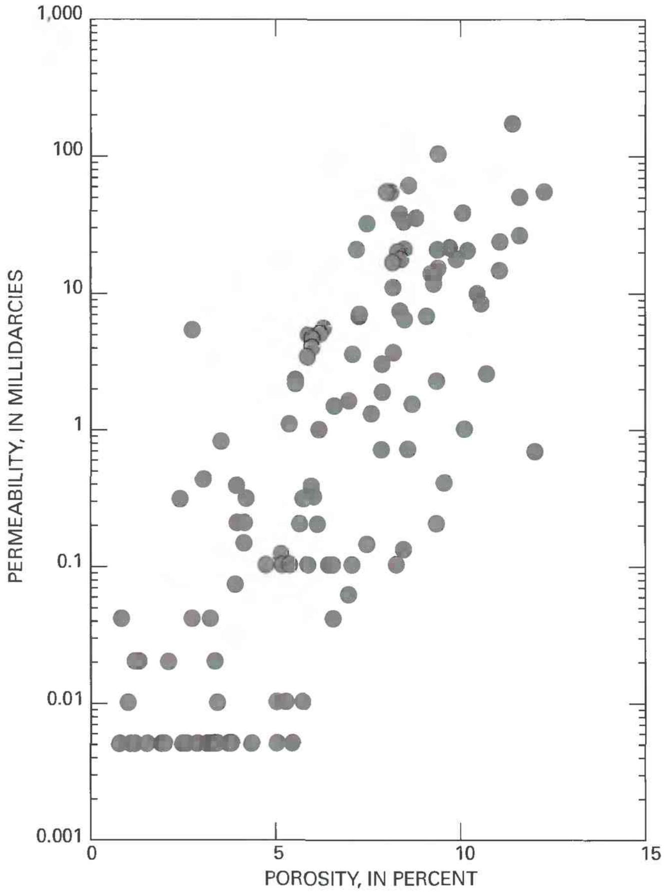


FIGURE 36.—Relation of porosity to pore-scale permeability in sandstone samples from the Elbert Formation.

fig. 43). On the basis of limestone to dolomite ratios, median values of porosity for limestone and dolomite, and scattered measurements, unit-averaged porosity in the study area is estimated to range regionally from less than 1 to 11 percent (pl. 2).

Most samples of limestone and dolomite in this hydrogeologic unit, regardless of texture, have little pore-scale permeability, although some samples can be very permeable. The pore-scale permeability of limestone ranges from less than 0.01 to 940 md, with a median value of less than 0.01 md. The pore-scale permeability of dolomite ranged from less than 0.01 to 673 md, with a median value of 0.10 md. Median values of pore-scale permeability were less than 0.01 md for all varieties of limestone, whereas several varieties of dolomite, including anhydritic or cherty, crystalline, and vuggy dolomite, had median values of pore-scale permeability of 0.01 md or larger (table 8). On the average, dolomite appears to be at least 10 times more permeable than limestone at pore scale.

In neither dolomite nor limestone does there appear to be any correlation between porosity and pore-scale permeability (fig. 44). This is attributed to the influences of secondary mineralization, fracturing, and solution on permeability. Carbonate rocks with large porosity and small pore-scale permeability probably have large unconnected vugs or may have pores partially filled with secondary calcite, silica, or anhydrite. Carbonate rocks with small porosity and relatively large pore-scale permeability probably contain fractures (see Thayer, 1983, p. 19–20).

Permeability values determined from field tests span the range from unfractured carbonate rocks with small permeability to fractured rocks with large permeability. In 136 drill-stem tests, local-scale permeability values ranged from 0.02 to 1,800 md, with a median value of 5.6 md. Local-scale permeability values of limestone ranged from 0.30 to 42 md, with a median value of 2.9 md. Local-scale permeability values of dolomite ranged from 0.027 to 540 md, with a median value of 12 md. Therefore, both the laboratory and drill-stem-test data indicate that dolomite in this hydrogeologic unit is more permeable than limestone.

HYDRAULIC CONDUCTIVITY, TRANSMISSIVITY, AND STORATIVITY

Hydraulic conductivity in intervals of the Redwall-Leadville zone ranges from small to large. In 172 tests of various kinds (pl. 2), the hydraulic conductivity of limestone and dolomite intervals ranged from 0.00005 to 200 ft/d, with a median value of 0.023 ft/d. Median values for limestone and dolomite were not significantly different in these tests (fig. 45). Pressure-injection tests by the Bureau of Reclamation at a damsite in the Uinta Mountains indicate that the hydraulic

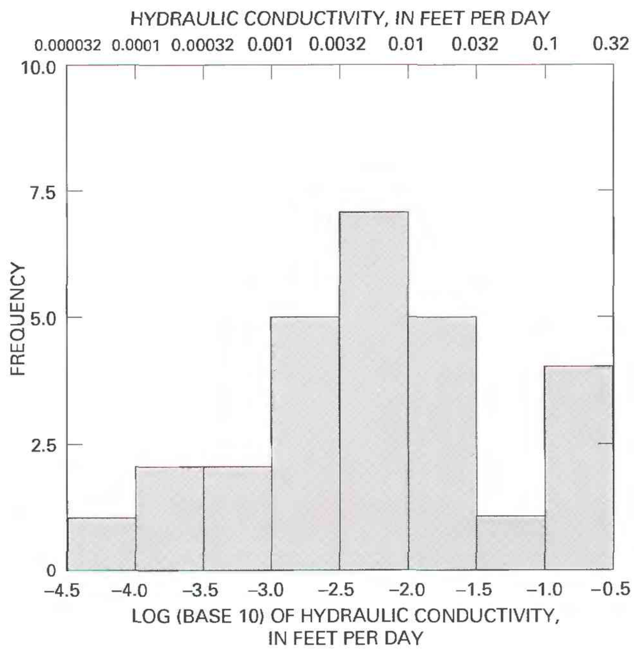


FIGURE 37.—Frequency distribution of hydraulic conductivity in the Elbert Formation.

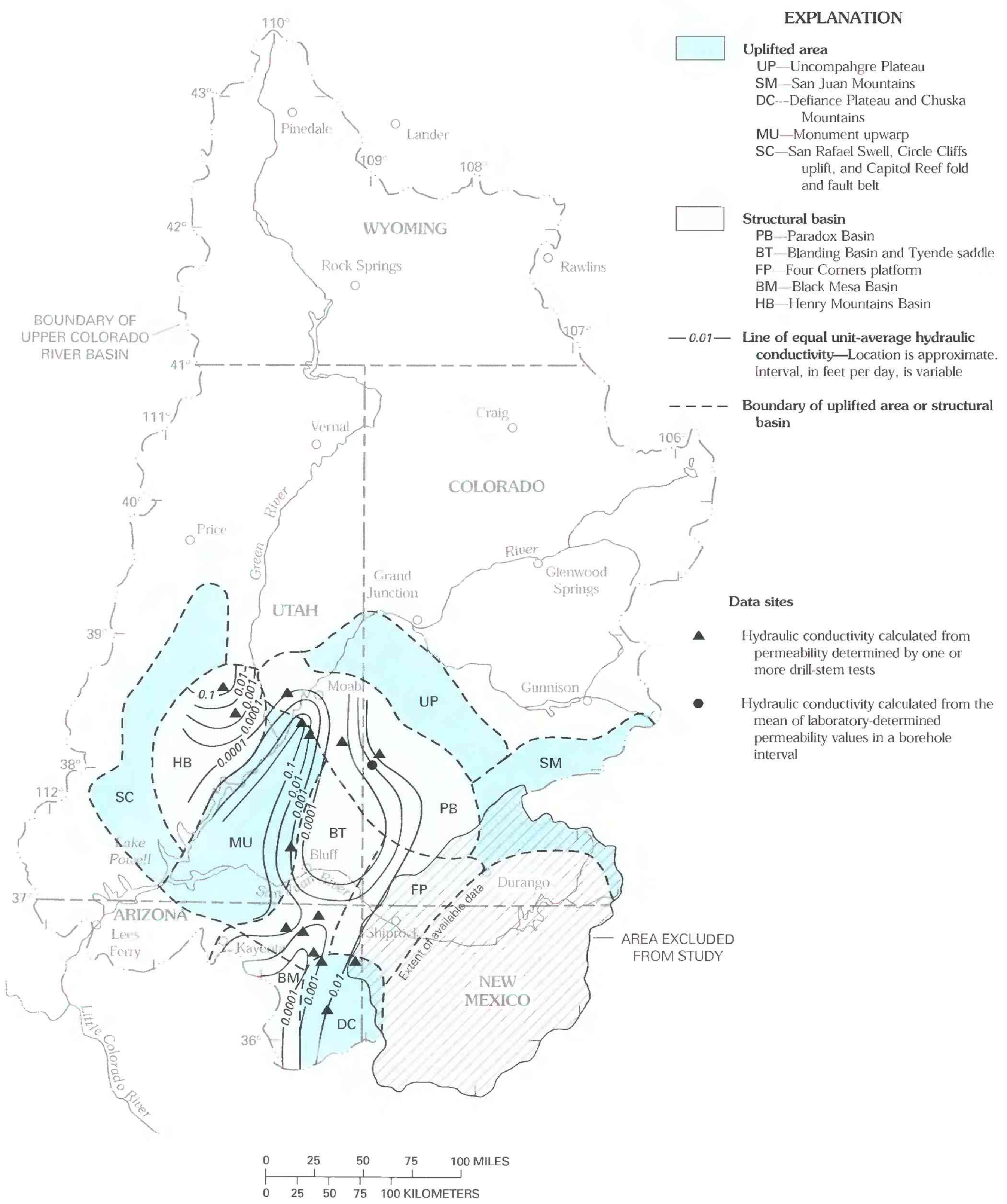


FIGURE 38.—Estimated areal distribution of unit-averaged hydraulic conductivity in the Elbert Formation.

FIGURE 39.—Estimated distribution of composite transmissivity in the Elbert Formation.

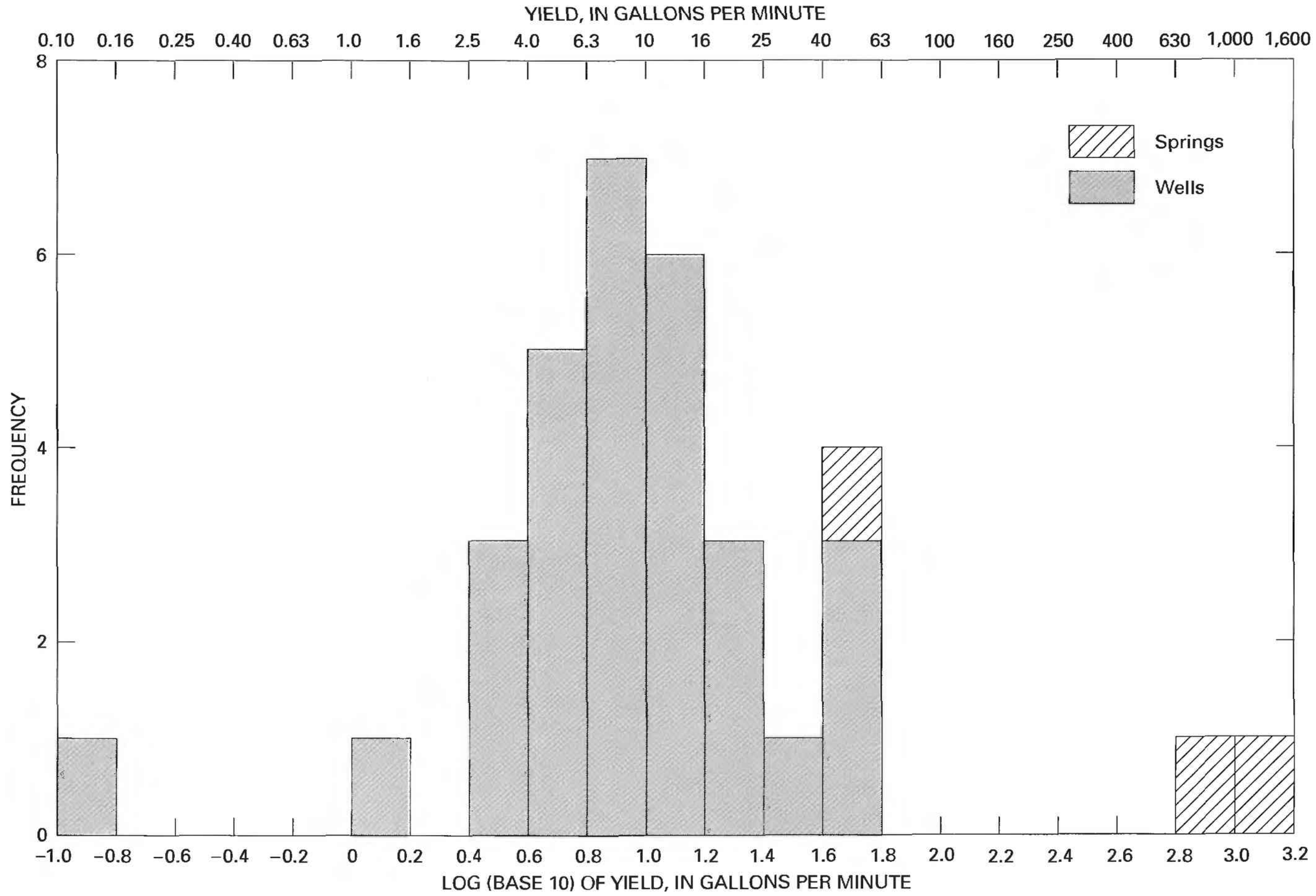


FIGURE 40.—Frequency distribution of yields from the Elbert-Parting confining unit.

conductivity of carbonate rocks appears to be related more to bedding thickness, joint spacing, brecciation, openness of fractures, vugginess, and the presence of chert nodules and beds than to composition (fig. 46). Reflecting the greater amount of open fractures and solution channels in uplifted areas, unit-averaged hydraulic conductivity increased from 0.00005 ft/d in structural basins to 200 ft/d in uplifted areas (pl. 2).

Aquifer tests at Ouray, Colo., indicate how dependent transmissivity in the Redwall-Leadville zone is on fractures and solution channels. Drawdown with time caused by four successive increments of discharge in a step-drawdown pumping test of well OX–3 (fig. 5) indicated values of transmissivity ranging from 130 to 280 ft^2/d, with an average value of about 200 ft^2/d. At completion, the flow into well OX–3 was measured as 120 gal/min. Located just 245 ft from well OX–3, well OX–5 (NMB44–07–31cbd$_4$) flowed at a rate of only 2 gal/min. As illustrated in figure 47, head recovery in well OX–5 during an airlift test of the Leadville Limestone on July 5, 1987, indicated the following:

$$\beta = Tt/r_c^2 = 1.0 \tag{23}$$

$$\alpha = \frac{r_w^2 S}{r_c^2} = 0.001 \tag{24}$$

$$T = \frac{\beta r_c^2}{t} = \frac{1.0 \times 0.29\text{ft}^2}{3{,}600 \text{ sec} \times 1\text{day}/86{,}400 \text{ sec}}$$

$$= 2\text{ft}^2/\text{d}$$

where

T = transmissivity, in feet squared per day;
t = time since slug removed, in seconds;
r_c = radius of casing, in feet;
r_w = effective well radius, in feet; and
S = storativity, dimensionless.

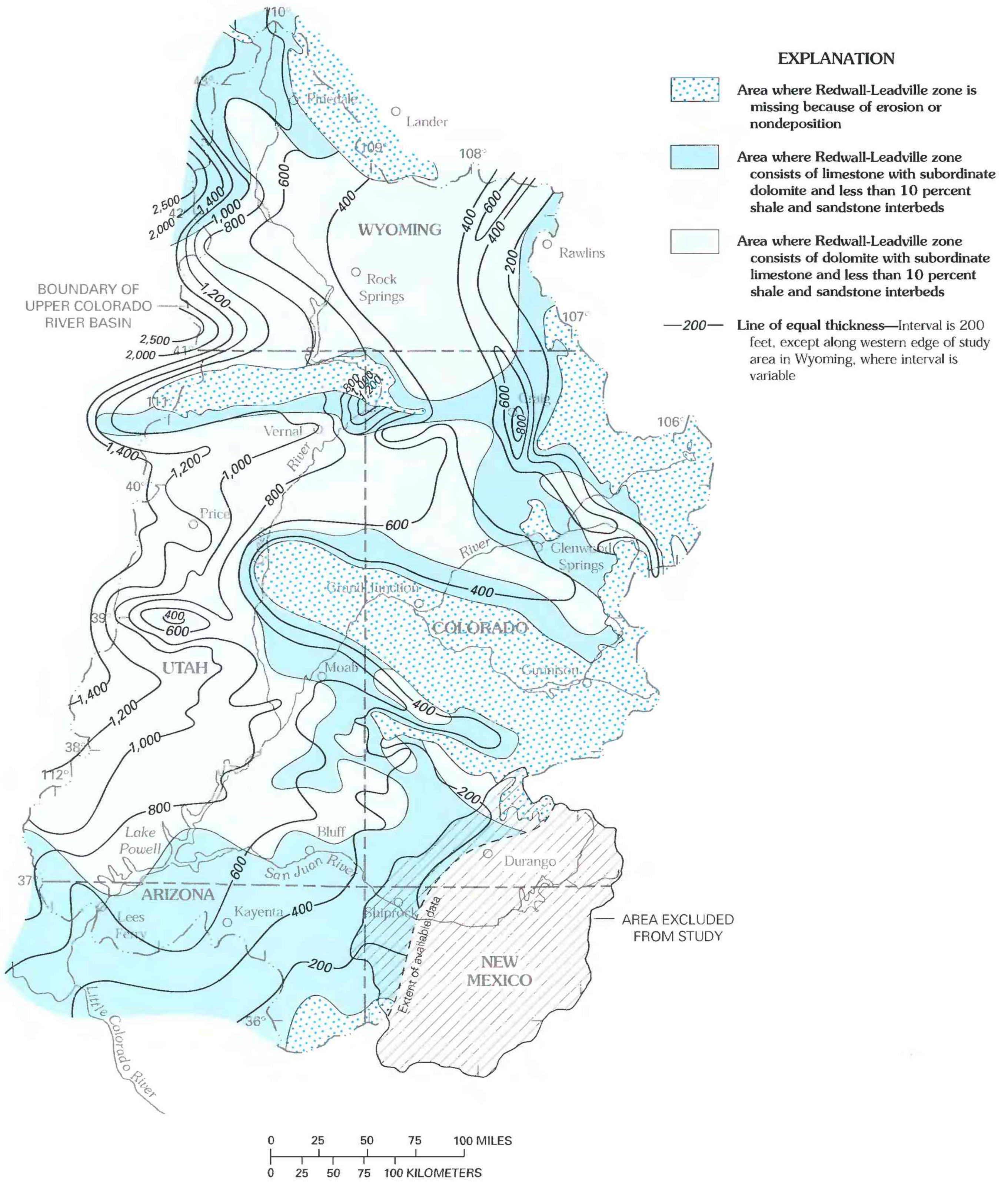


FIGURE 41.—Thickness and lithology of the Redwall-Leadville zone of the Madison aquifer. (Modified from Geldon, in press, pl. 11.)

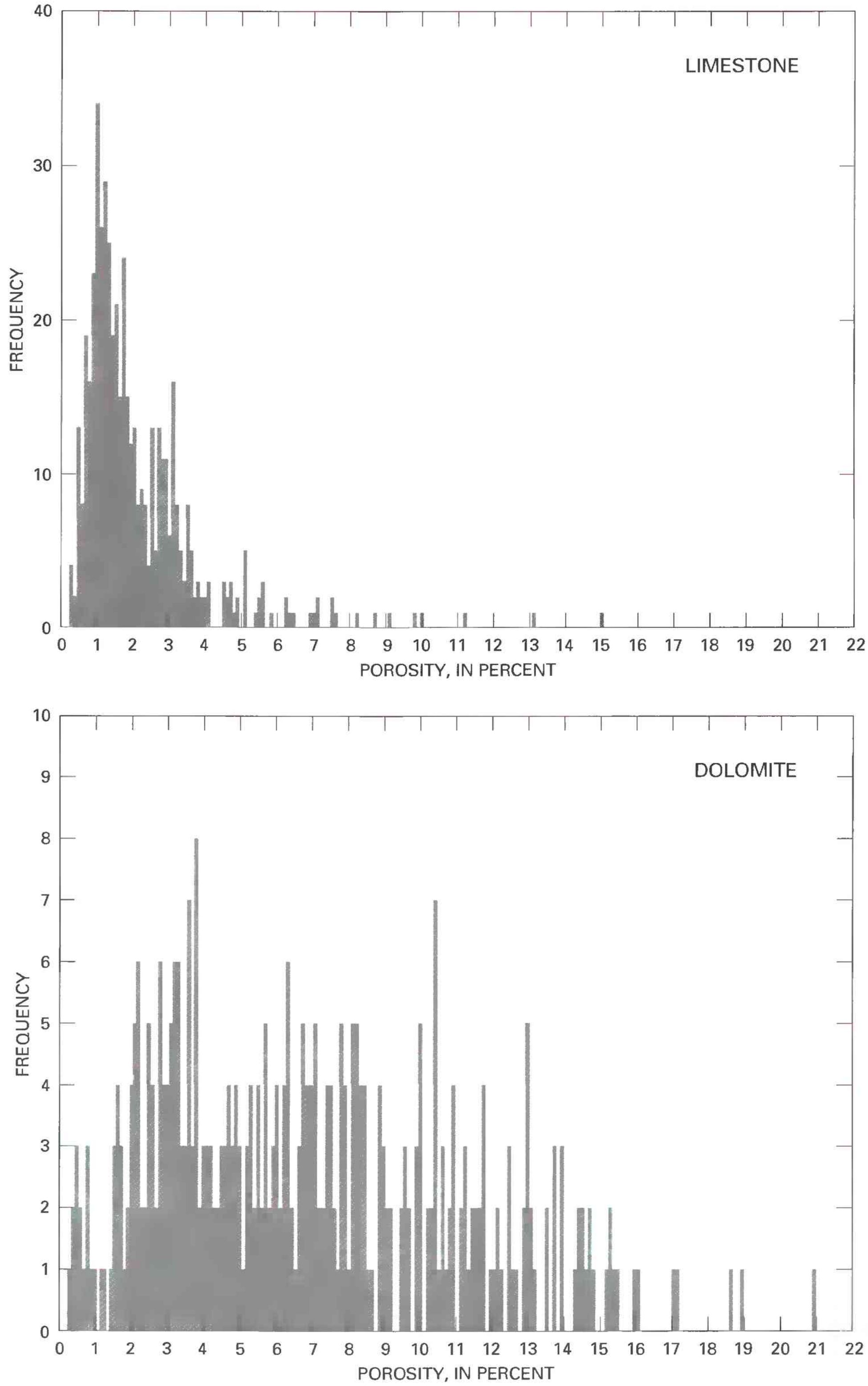


FIGURE 42.—Frequency distribution of porosity in samples of limestone and dolomite from the Redwall-Leadville zone of the Madison aquifer.

TABLE 8.—*Porosity and pore-scale permeability statistics for the Redwall-Leadville zone of the Madison aquifer*

[<, less than]

Rock type	Porosity (percent)			Number of observations	Pore-scale permeability (millidarcies)			Number of observations
	Minimum	Maximum	Median		Minimum	Maximum	Median	
Limestone								
Oolitic	1.2	3.1	2.4	10	<0.01	<0.01	<0.01	10
Shaly	.3	4.1	.8	48	<.01	20.0	<.01	48
Fossiliferous	.4	5.1	2.6	149	<.01	3.1	<.01	149
Anhydritic or cherty	1.2	3.9	1.9	10	<.01	356	<.01	10
Algal	1.0	2.8	1.6	16	<.01	389	<.01	16
Crystalline	.3	10.0	1.2	243	<.01	940	<.01	244
Sucrosic	1.4	3.2	1.8	8	<.01	<.01	<.01	8
All	.3	15.0	1.7	513	<.01	940	<.01	513
Dolomite								
Shaly	1.6	10.0	3.3	49	<.01	5.9	<.01	49
Fossiliferous	3.2	15.9	5.7	8	<.01	24	<.01	8
Anhydritic or cherty	.8	17.1	8.3	112	<.01	13.3	<.01	112
Vuggy	2.0	15.4	10.2	17	<.01	13.3	2.4	17
Fine-grained	.7	4.9	1.9	8	<.01	<.01	<.01	8
Crystalline	.5	21.3	7.8	175	<.01	335	.05	175
All	.3	21.6	6.5	435	<.01	673	.10	435
Shale	1.5	4.0	2.8	2	<.01	<.01	<.01	2
Sandstone[1]	.9	21	13	6	.03	119	24	6
Chert[1]	.6	19	11	4	.28	.72	.45	3
Anhydrite[1]	.3	1.3	.6	10	.01	.06	.03	4

[1]Samples from the Powder River Basin and vicinity in northeastern Wyoming and south-central Montana (from Thayer, 1983, p. 7). Values in median columns are the means of analyses; the median values were not reported.

The substantial differences in transmissivity and yield determined in the 245 feet between wells OX–3 and OX–5 indicate that the Leadville Limestone at Ouray is extremely heterogeneous. Hydrologic and geologic reports prepared by consultants for the City of Ouray (David Vince, City of Ouray, written commun., 1988) indicate that this heterogeneity probably is related to a site's position with respect to the Ouray or Portland Creek faults, splays from these faults, or fracture zones related to the faulting. One consultant concluded that all 12 of the hot and warm springs mapped at Ouray issue from faults, joints, or bedding planes, and none appear to be related to folds, breccia zones, or formational contacts.

The maximum value of transmissivity for the Redwall-Leadville zone of the Madison aquifer in the UCRB was determined during the RASA study from a flowing-well test of the Leadville Limestone at Glenwood Springs, Colo. (see Geldon, 1989c, for test description and interpretation). This test was done in November 1984. In the test area, the Redwall-Leadville zone consists of the Dyer Dolomite and Leadville Limestone, which are transected by faults and, consequently, are highly fractured. Caves in the Leadville Limestone, some of which contain hot water and steam, are visible near the test site. During the aquifer test, a geothermal well (SC06–89–09bba) was allowed to flow for 4 days, discharging a volume of 10.5 million gallons of water. This discharge decreased head in the production well by about 8 ft, lowered head in an observation well about 0.8 mi away by about 1 ft, and reduced discharges from springs within a radius of 1,100 ft. From analyses of drawdown and recovery data from the production and observation wells, a median transmissivity value of 47,000 ft^2/d was calculated.

As shown on plate 3, the composite transmissivity of the Redwall-Leadville zone is estimated to range regionally from 0.01 to 47,000 ft^2/d. These estimates are based on three aquifer tests and the regional distributions of the thickness and unit-averaged hydraulic conductivity of the Redwall-Leadville zone. For comparison, three aquifer tests of the Madison Limestone in

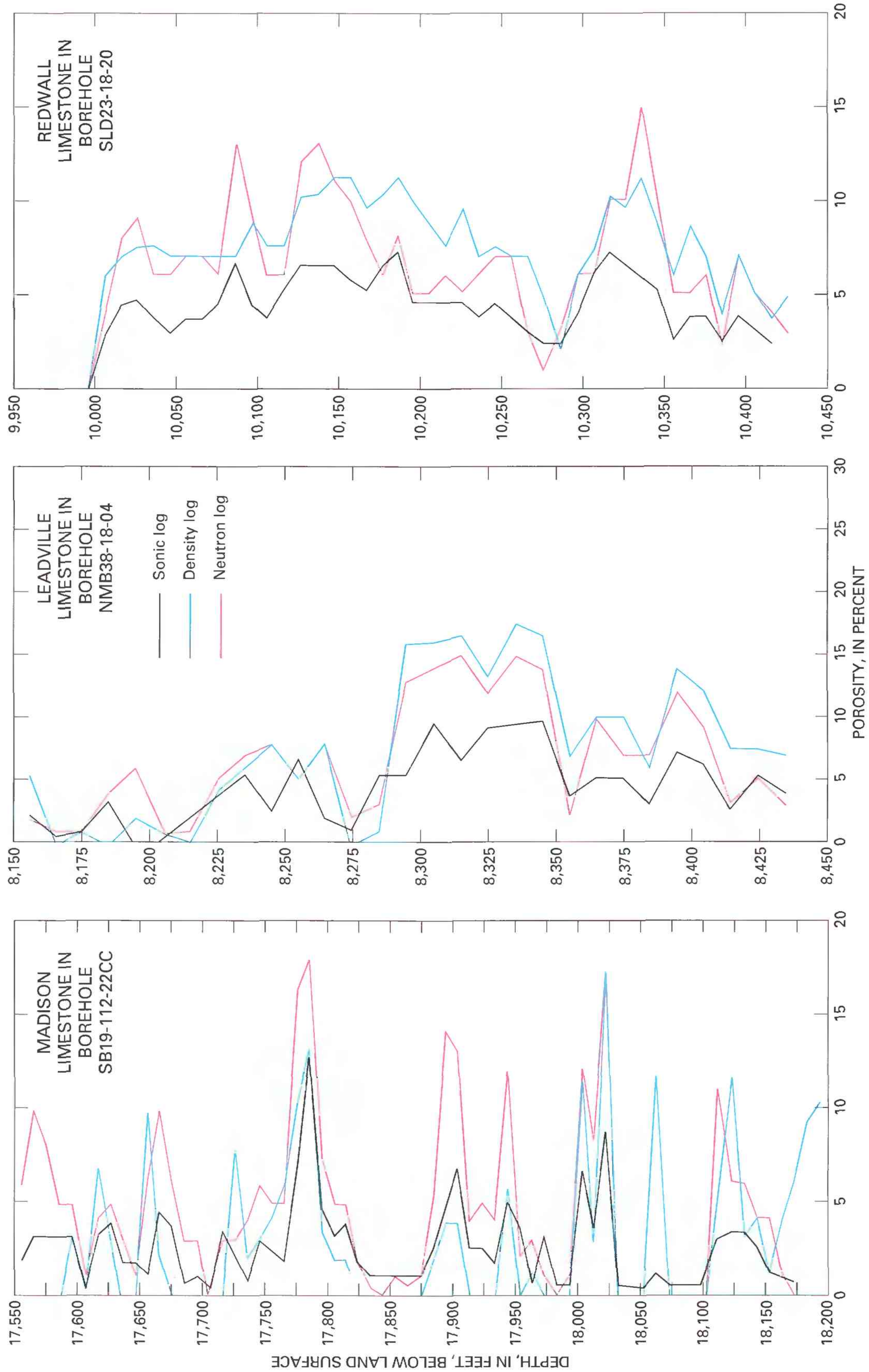


FIGURE 43.—Relation of geophysically determined porosity to depth below land surface in component geologic units of the Redwall-Leadville zone of the Madison aquifer.

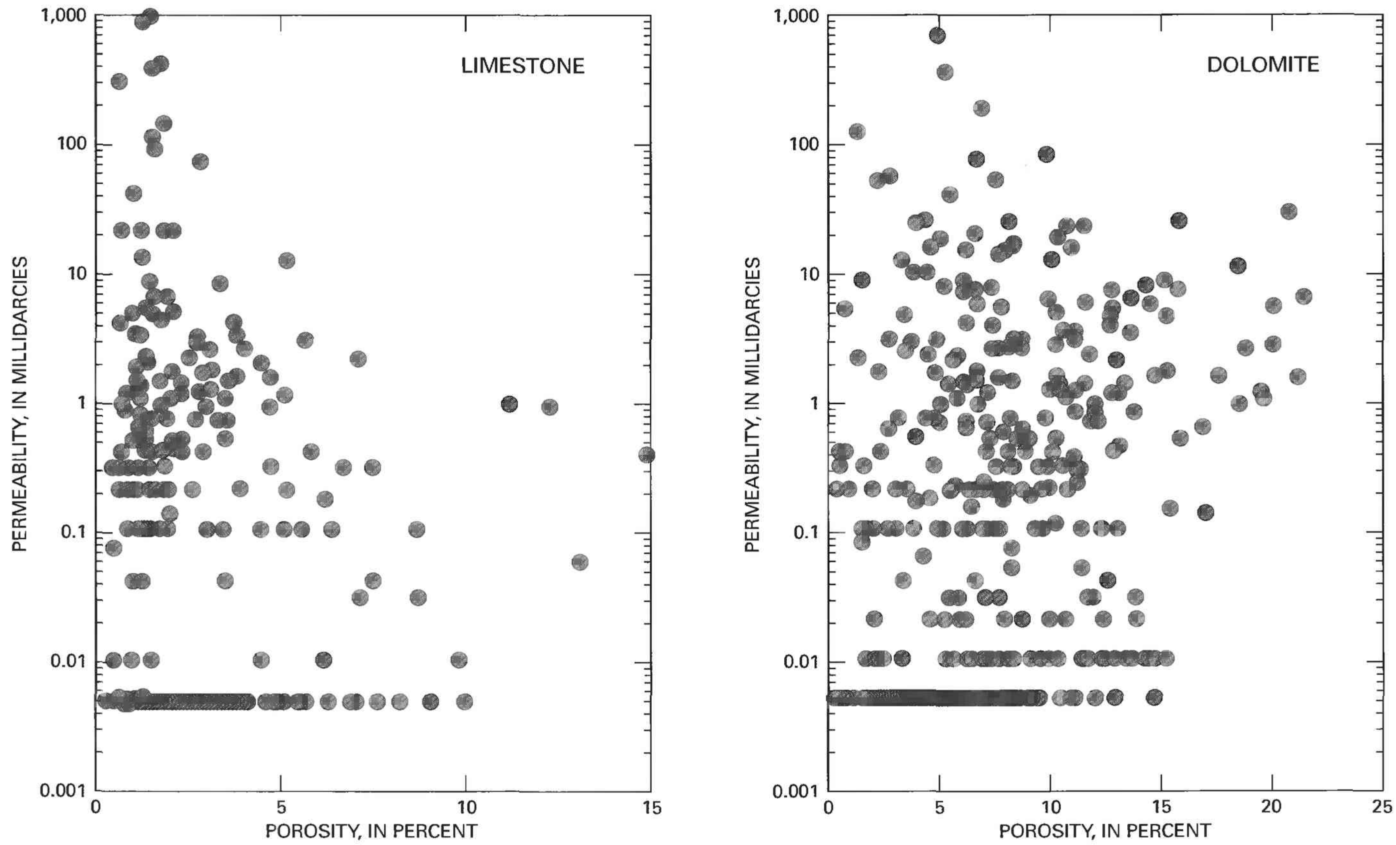


FIGURE 44.—Relation of porosity to pore-scale permeability in samples of limestone and dolomite from the Redwall-Leadville zone of the Madison aquifer.

the Bighorn Basin of Wyoming indicated transmissivity values ranging from 72 to 1,900 ft^2/d (Cooley, 1985, p. 39). Downey (1984, p. 22–23) estimated the transmissivity of the Lodgepole and Mission Canyon Limestones in the Powder River Basin of northeastern Wyoming and south-central Montana to range from less than 250 to more than 4,000 ft^2/d using the three-dimensional finite-difference model of Trescott (1975). Drill-stem tests of the Madison Group in Montana indicated transmissivity values ranging from 0.043 to 5,400 ft^2/d (Konikow, 1976, p. 10; Downey, 1984, p. 23). Step-drawdown pumping tests at one site in Montana indicated transmissivity values of 32,000 to 45,000 ft^2/d. Collectively, the data from Wyoming and Montana indicate a transmissivity range of 0.043 to 45,000 ft^2/d for rocks equivalent to those of the Redwall-Leadville zone, which supports the estimated range in composite transmissivity for the Redwall-Leadville zone in the UCRB.

The aquifer test at Glenwood Springs indicated a storativity of 0.0005. Dividing this value by the thickness of the Redwall-Leadville zone in the test area (280 ft) gives a specific storage of 0.000002 ft^{-1}. If this specific storage applies for the entire UCRB, then the storativity where the Redwall-Leadville zone is at least 100 ft thick would range from 0.0002 to 0.005.

YIELDS FROM WELLS AND SPRINGS

Yields from the Redwall-Leadville zone of the Madison aquifer to wells and springs are small to moderate in most areas of the UCRB, but they can be very large in uplifted areas near faults (pl. 3). Yields from 227 drill-stem tests, flowing wells, and springs ranged from less than 1 to 45,000 gal/min, with a median value of 24 gal/min (fig. 48). Yields from wells in basins typically are less than 50 gal/min, whereas many wells and springs in or on the margins of uplifted areas discharge at rates of several hundred to several thousand gallons per minute (see, for example, Boettcher, 1972; Hampton, 1974; Lines and Glass, 1975; Hood and others, 1976; Sumsion, 1976). Flowing wells with large discharges include the Redstone 21–9 well (SC06–89–09bba) at Glenwood Springs, Colo., (1,500 to 2,300 gal/min), and the Paradise no. 3 well (SC02–84–02bc) at McCoy, Colo. (3,200 to 9,400 gal/min). Large springs include Big Spring (UB01–08–17cbb) in the Uinta Mountains (2,240 to 6,940 gal/min), Hogsback Spring (SB26–114–01bac) in the Overthrust Belt (5,500 gal/min), and the Yampa Hot Spring (SC06–89–09ada) at Glenwood Springs, Colo. (2,950 gal/min). Collectively, 18 hot springs and seepage areas at Glenwood Springs discharge at a rate of 4,300 gal/min (Geldon, 1989c). Geothermal wells and springs at Ouray, Colo., discharge at

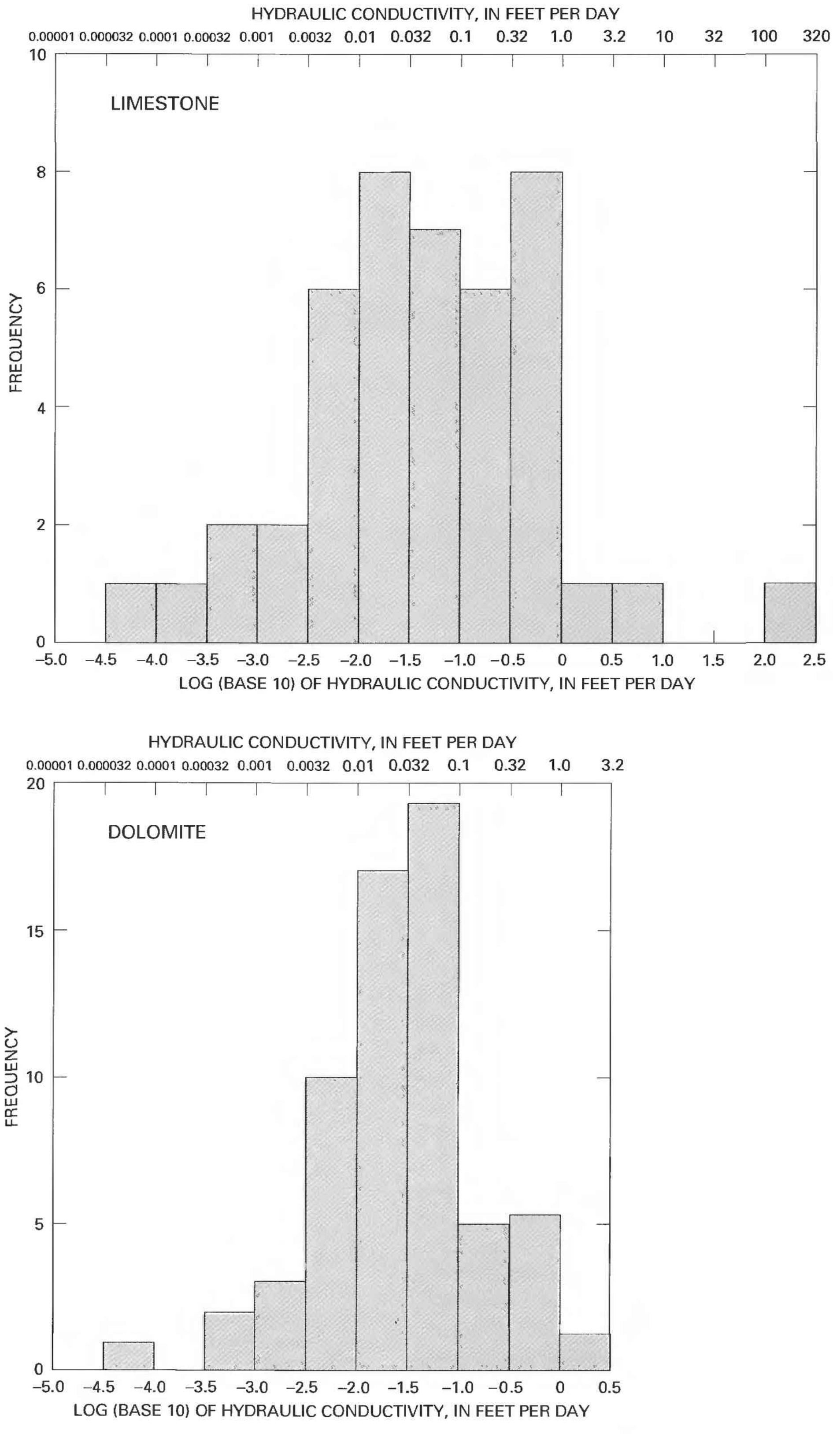


FIGURE 45.—Frequency distribution of hydraulic conductivity in limestone and dolomite intervals of the Redwall-Leadville zone of the Madison aquifer.

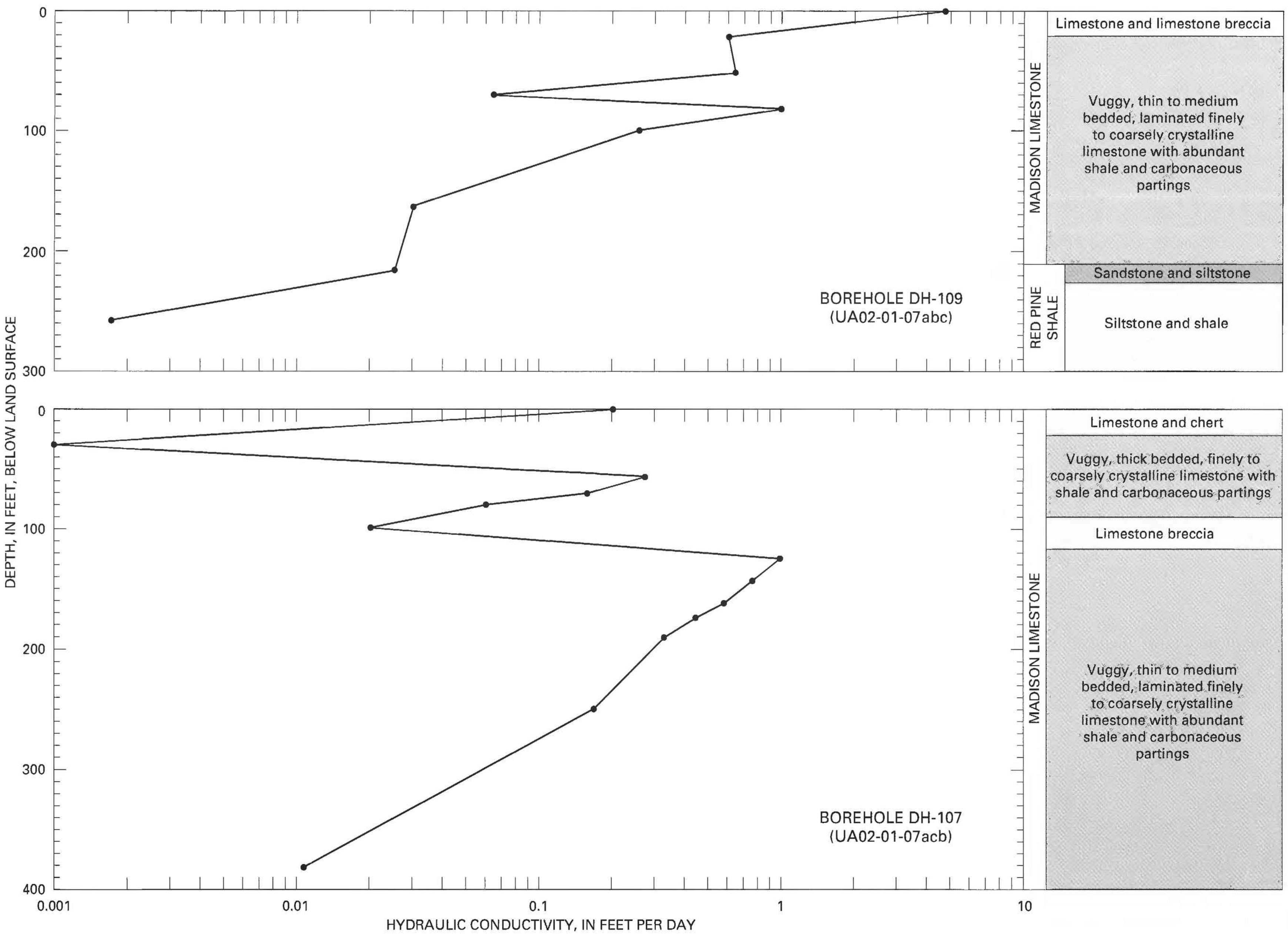


FIGURE 46.—Relation of the hydraulic conductivity of the Madison Limestone to depth below land surface in boreholes at the Whiterocks River damsite, Utah (data furnished by Bureau of Reclamation, written commun., 1983–85).

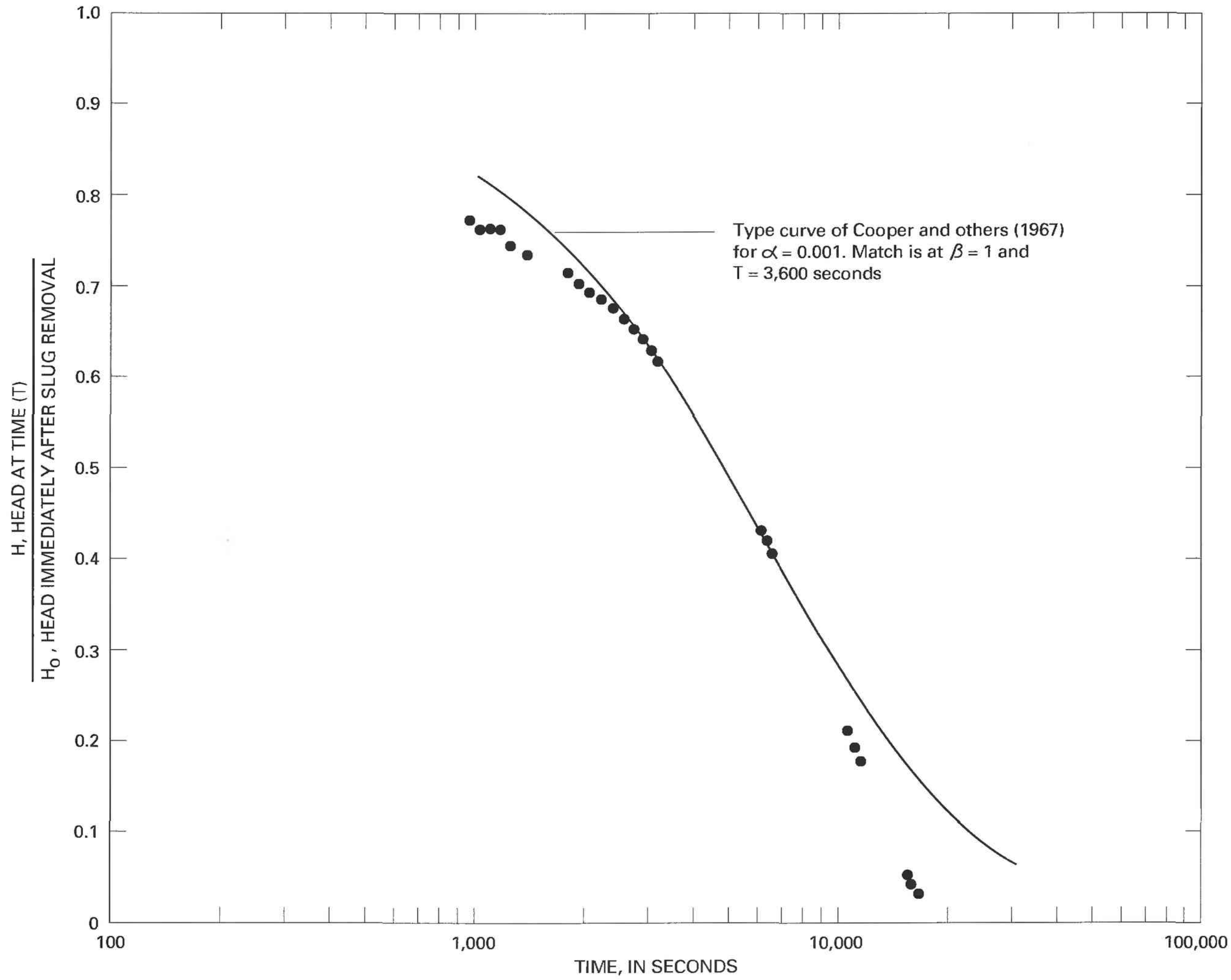


FIGURE 47.—Head recovery in well OX–5 (NMB44–07–31cbd$_4$) during "airlift test" of Leadville Limestone at Ouray, Colorado, July 15, 1987 (data from David Vince, City of Ouray, written commun., 1988).

average rates ranging from 2 to 340 gal/min, and, collectively, seven warm and hot springs at Ouray discharge from the Leadville Limestone or overlying alluvium at an average rate of 672 gal/min (David Vince, City of Ouray, written commun., 1988).

The largest discharges from the Redwall-Leadville zone result from unusual hydrological conditions. Pole Creek Spring (UB03–02–34d), with a discharge ranging from 900 to 11,000 gal/min, is the outlet for Pole Creek Sink (Maxwell and others, 1971, p. 17). Blue Spring (GA32–07–31b) and two other springs in the canyon of the Little Colorado River with discharges ranging from 11,000 to 45,000 gal/min (Cooley, 1976) are outlets for the entire Four Corners aquifer system and probably the Canyonlands aquifer, as well, for all of the area between the Uncompahgre Plateau and Mogollon Rim (see pl. 1 for locations of these areas). Springs issuing downstream from sinks or in subregional discharge areas result more from unique topographic situations and the flow-system configuration than from the transmissivity of the Redwall-Leadville zone and, consequently, are considered atypical of the Redwall-Leadville zone in the UCRB.

DARWIN-HUMBUG ZONE OF THE MADISON AQUIFER

The Darwin-Humbug zone of the Madison aquifer is present only in and near the Uinta Mountains and Uinta Basin and on the flanks of uplifts bordering the greater Green River Basin (the Overthrust Belt, Gros Ventre Range, and Wind River Mountains). It consists of the Humbug Formation, the Bull Ridge Member of the Madison Limestone, the upper part of the Mission Canyon Limestone, and the Darwin Sandstone Member of the Amsden Formation (table 1). Component geologic units are Early to Late Mississippian in age.

Little has been written about the water-supply capabilities of the geologic units that compose the Darwin-Humbug zone, and few data have been collected to quantify hydrologic properties of this zone. In the Uinta Mountains, the Humbug Formation contains fractures, solution channels, and breccia zones that can transmit water. A combination of these features allows the flow of Little Brush Creek to disappear into a cavern in the Humbug Formation, travel through the Humbug Formation to the Madison Limestone, and emerge as a spring in Brush Creek (Maxwell and others, 1971). The Darwin Sandstone Member of the Amsden Formation probably transmits water readily in the UCRB because it yields small quantities of water to wells and springs in the Bighorn Basin (Cooley, 1985, p. 8). Based on their lithology, the Bull Ridge Member of the Madison Limestone and the upper

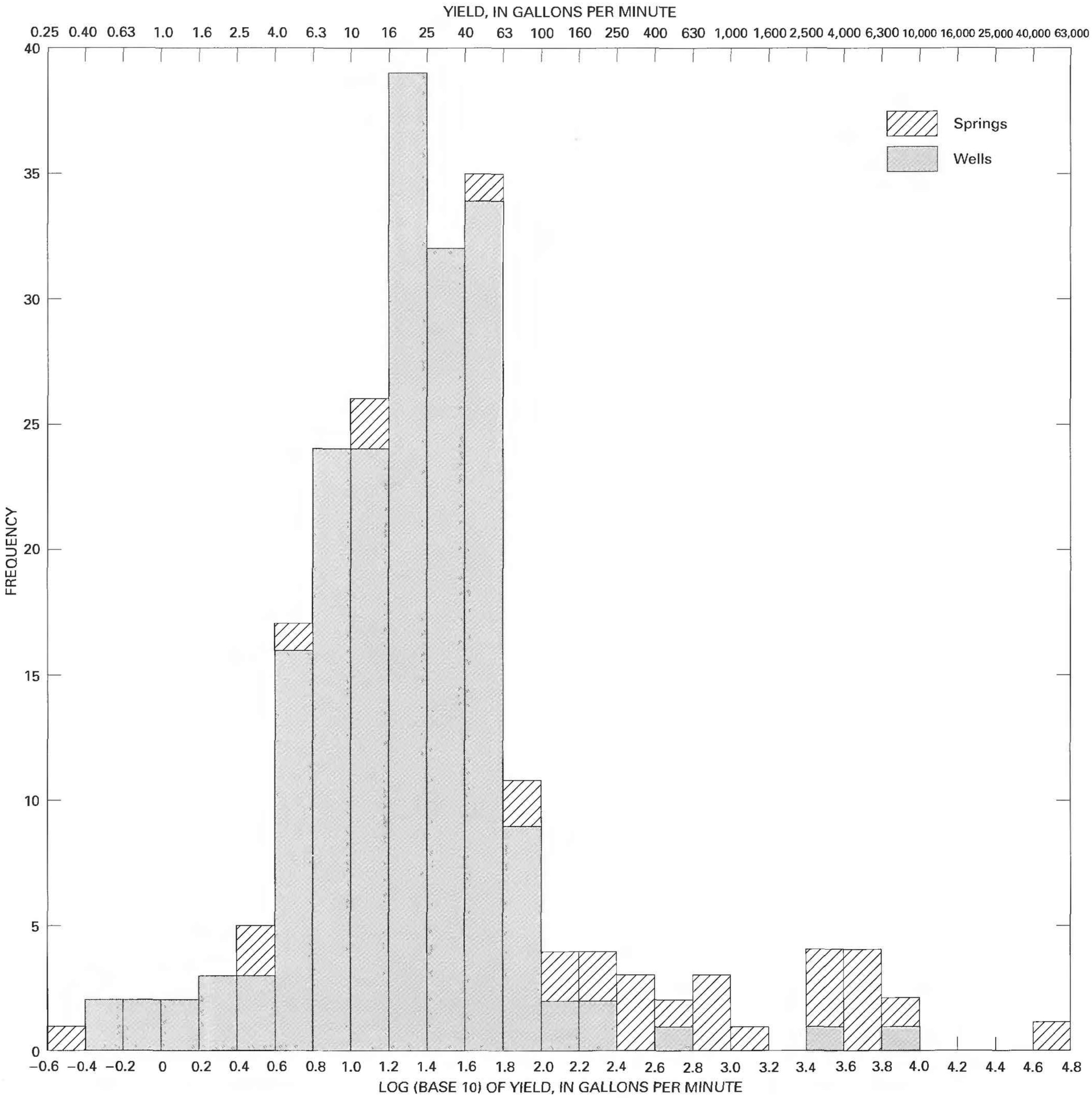


FIGURE 48.—Frequency distribution of yields from the Redwall-Leadville zone of the Madison aquifer.

Mission Canyon Limestone probably have hydrologic properties similar to the Humbug Formation and Darwin Sandstone Member. Where fractures, solution channels, breccia zones, or loosely cemented sandstone interbeds are absent, the Darwin-Humbug zone probably does not transmit much water. For this reason and because of its limited distribution, the Darwin-Humbug zone can be expected to function as an aquifer subregionally in the UCRB.

THICKNESS AND LITHOLOGY

The thickness of the Darwin-Humbug zone ranges from 0 to more than 600 ft (fig. 49). The hydrogeologic unit consists of variable proportions of limestone, dolomite, sandstone, shale, solution breccia, and gypsum. Carbonate rocks account for more than 75 percent of the Darwin-Humbug zone in the centers of depositional areas. Sandstone layers increase in abundance away from depositional centers as carbonate layers decrease. Carbonate layers typically are brecciated, and sandstone layers are brecciated in the western part of the UCRB. Gypsum layers are abundant at the northern end of the Hoback Range in Wyoming. Contacts with Mississippian and Pennsylvanian formations that comprise the overlying Four Corners confining unit generally are conformable.

POROSITY, PERMEABILITY, AND HYDRAULIC CONDUCTIVITY

Regional variations in porosity cannot be established with certainty from the limited data that were obtained during this study. In 13 samples of sandstone from the Humbug Formation, porosity ranged from 1.4 to 16 percent, with a median value of 8.6 percent. It is estimated that the Darwin Sandstone Member of the Amsden Formation and sandstone layers in the upper part of the Mission Canyon Limestone and Bull Ridge Member of the Madison Limestone have a similar range in porosity. The porosity of carbonate layers probably is similar to that of carbonate rocks in the Redwall-Leadville zone of the Madison aquifer (see table 8) into which the Darwin-Humbug zone grades.

The small amount of permeability and hydraulic-conductivity data for the Darwin-Humbug zone that was available was consistently in the moderate range. Samples of sandstone from the Humbug Formation obtained from a borehole near Jensen, Utah (SLD05–23–18cca), had pore-scale permeability values ranging from 0.08 to 60 md, with a median value of 7.5 md (which is equivalent to a hydraulic-conductivity value of 0.011 ft/d). A drill-stem test in a borehole (SB08–99–17) near Maybell, Colo., indicated a local-scale permeability value of 292 md and a hydraulic-conductivity value of 0.71 ft/d for interbedded dolomite, sandstone, and shale in the Humbug Formation. These data are too limited to be representative of regional variations in permeability and hydraulic conductivity within the Darwin-Humbug zone that might be expected from the lithology of the zone.

TRANSMISSIVITY

The drill-stem test near Maybell indicated a transmissivity of 110 ft/d for the Humbug Formation, assuming that the tested interval is representative of the entire formation at the test site. The drill-stem test was done on the Axial Basin Arch, which, like other uplifted areas, contains a fracture system that enhances secondary permeability and transmissivity. It is estimated that transmissivity values in structural basins are smaller than the value indicated by the drill-stem test on the Axial Basin Arch. Conversely, transmissivity values in areas with well-developed networks of solution channels and fractures, such as the southeastern Uinta Mountains, are estimated to be larger than the value indicated by the Maybell drill-stem test. On the basis of the rock types present in the Darwin-Humbug zone and the existence of fracture-solution channel networks within this zone, the Darwin-Humbug zone is believed to be almost as transmissive as the Redwall-Leadville zone in places, and its transmissivity tentatively is estimated to range from 0.01 to 1,000 ft^2/d.

YIELDS FROM WELLS AND SPRINGS

Yields of water from the Darwin-Humbug zone to wells typically do not exceed 50 gal/min. In four drill-stem tests of the Humbug Formation and the Darwin Sandstone Member of the Amsden Formation, yields ranged from 7.2 to 34 gal/min, with a median value of 8.7 gal/min. However, a spring issuing from the Humbug Formation in the Uinta Mountains (UB02–07–36aba) has an estimated discharge of 900 gal/min. According to Hood (1976, p. 11), such large discharges are possible from the Humbug Formation wherever a karst topography has developed. By analogy, similarly large discharges to wells and springs from the upper part of the Mission Canyon Limestone and the Bull Ridge Member of the Madison Limestone may be possible where karst topography has developed in the Overthrust Belt, Gros Ventre Range, and Wind River Mountains.

HYDROLOGIC PROPERTIES OF THE FOUR CORNERS CONFINING UNIT

Overlying the Four Corners aquifer system throughout most of the UCRB is a thick stratigraphic sequence composed mostly of shale, anhydrite, gypsum, halite, and carbonate rocks that generally inhibits ground-water movement between overlying and underlying aquifers. These rocks comprise the Four Corners confining unit, which is absent only in Wyoming east of the Overthrust Belt and north of the Rock Springs Uplift, in the central UCRB on and adjacent to the Uncompahgre Plateau, and along the southwestern edge of the UCRB in and adjacent to the High

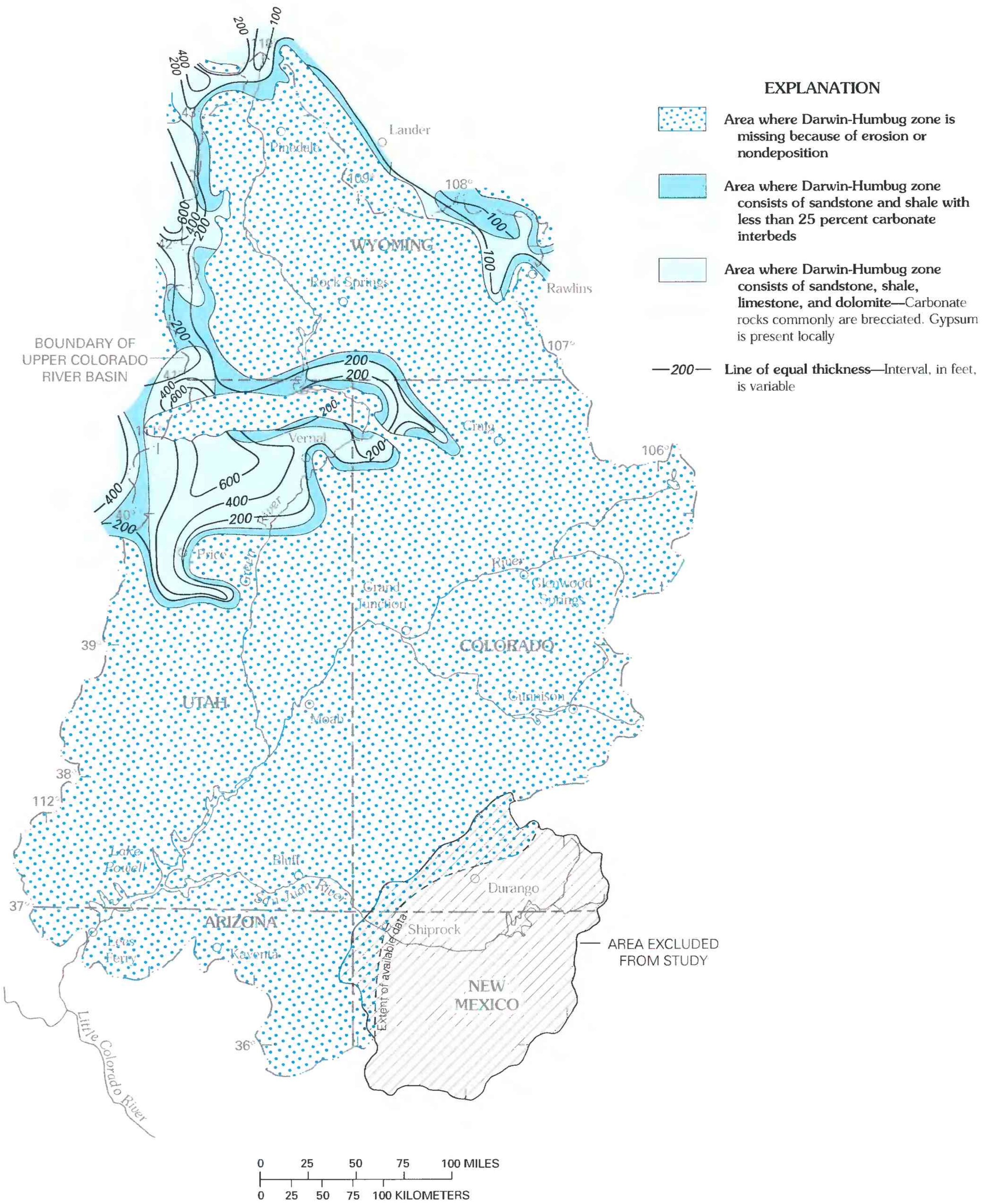


FIGURE 49.—Thickness and lithology of the Darwin-Humbug zone of the Madison aquifer. (Modified from Geldon, in press, pl. 12.)

Plateaus region. The Four Corners confining unit is divisible vertically into two subunits with different lithologic and hydrologic properties, the Belden-Molas and Paradox–Eagle Valley subunits.

BELDEN-MOLAS SUBUNIT

The Belden-Molas subunit of the Four Corners confining unit consists of the Horseshoe Shale Member of the Amsden Formation, Doughnut Shale, Surprise Canyon Formation, Watahomigi Formation, Molas Formation, lower member of the Hermosa Formation, and Belden Formation (table 1), which are Late Mississippian to Middle Pennsylvanian in age. Because component geologic units are mostly shale, the Belden-Molas subunit generally has negligible permeability. In the Paradox Basin, for example, drill-stem tests of the Molas Formation and lower member of the Hermosa Formation produce so little water that only the Paradox Member of the Hermosa Formation and, in some of the area, the Ignacio Quartzite are considered to be less permeable (Rush and others, 1982; Weir and others, 1983a; Whitfield and others, 1983). The Doughnut Shale in the Uinta Mountains (Sumsion, 1976, p. 23), Uinta Basin (Hood, 1976, p. 11), and San Rafael Swell (Hood and Patterson, 1984, p. 55), the Horseshoe Shale Member of the Amsden Formation in the Rawlins Uplift (Berry, 1960, p. 13) and Overthrust Belt (Lines and Glass, 1975), and the Supai Group (including the Watahomigi Formation) in the Grand Canyon (Metzger, 1961, p. 118) also yield little or no water. The Belden Formation generally produces little or no water, but spring and artesian-well discharges occur in the White River Plateau and Elk Mountains where faults allow movement of water into the formation from the Leadville Limestone (Barrett and Pearl, 1977, p. 85; Geldon, 1989c). On the basis of predominant characteristics, this hydrogeologic unit is a major impediment to ground-water movement throughout the UCRB and, thus, was classified as part of a confining unit in this investigation.

THICKNESS AND LITHOLOGY

The thickness of the Belden-Molas subunit ranges from 0 to more than 3,000 ft (fig. 50). Regionally, the hydrogeologic unit consists predominantly of dark gray, black, and red shale with subordinate limestone, dolomite, and sandstone and minor gypsum. In most areas, shale accounts for 50 to more than 75 percent of the Belden-Molas subunit. However, in the Washakie Basin and depositional centers south of the Uinta Mountains and Axial Basin Arch, limestone and dolomite layers can equal or exceed shale layers in abundance, and, locally, carbonate rocks account for more than 75 percent of the unit. Sandstone layers compose about one-third of the unit where it thins on the flanks of uplifted areas. Contacts with Upper Mississippian and Pennsylvanian geologic units generally are conformable to gradational.

POROSITY AND PERMEABILITY

In considering the hydrologic properties of the Belden-Molas subunit, quantitative assessment was hindered by the sparse distribution of available data, most of which is from the lower member of the Hermosa Formation. However, it was possible to estimate hydrologic properties using data from approximately synchronous geologic units containing different proportions of the same rock types. In this analysis, data from the Paradox and upper members of the Hermosa Formation, Round Valley Limestone, Amsden Formation, and Cutler Formation were used.

The available data indicate that hydrologic properties of the Belden-Molas subunit depend on both lithology and structural setting. In 17 samples of limestone and dolomite from the lower member of the Hermosa Formation, porosity ranged from 0.3 to 7.9 percent, with a median value of 2.1 percent (fig. 51). On the basis of 48 samples of shale from the Hermosa Formation (fig. 51), it is estimated that the porosity of shale in the Belden-Molas subunit ranges from less than 1 to 13 percent, with a median value of about 2 percent. Analyses of samples from the Molas, Hermosa, and Cutler Formations indicate that in the Belden-Molas subunit, the median porosity of quartz sandstone probably is between 6 and 7 percent, and the median porosity of arkosic (red) sandstone probably is between 10 and 11 percent. On the basis of lithologic composition and the above-cited values of median porosity, it is estimated that the unit-averaged porosity of the Belden-Molas subunit ranges regionally from less than 2 to about 6 percent (fig. 52).

The permeability of the Belden-Molas subunit varies from small to moderate. In 17 analyses of limestone and dolomite from the lower member of the Hermosa Formation, pore-scale permeability ranged from less than 0.01 to 45 md, with a median value of less than 0.01 md. On the basis of 48 analyses for the entire Hermosa Formation (fig. 53), the pore-scale permeability of shale in the Belden-Molas subunit is estimated to range from less than 0.0001 to 18 md, with a median value of less than 0.01 md. On the basis of 176 analyses of samples from the Molas and Hermosa Formations, the pore-scale permeability of sandstone in the Belden-Molas subunit is estimated to range from 0.0001 to 25 md, with a median value of about 0.01 md. Because the data are limited, it is impossible to state conclusively whether pore-scale permeability in the Belden-Molas subunit depends on porosity, but a relationship appears to exist for dolomite (table 3).

Values of local-scale permeability in the Belden-Molas subunit reflect the influence of fracturing and other secondary openings and are larger, on the average, than values of pore-scale permeability. In 15 drill-stem tests, local-scale permeability

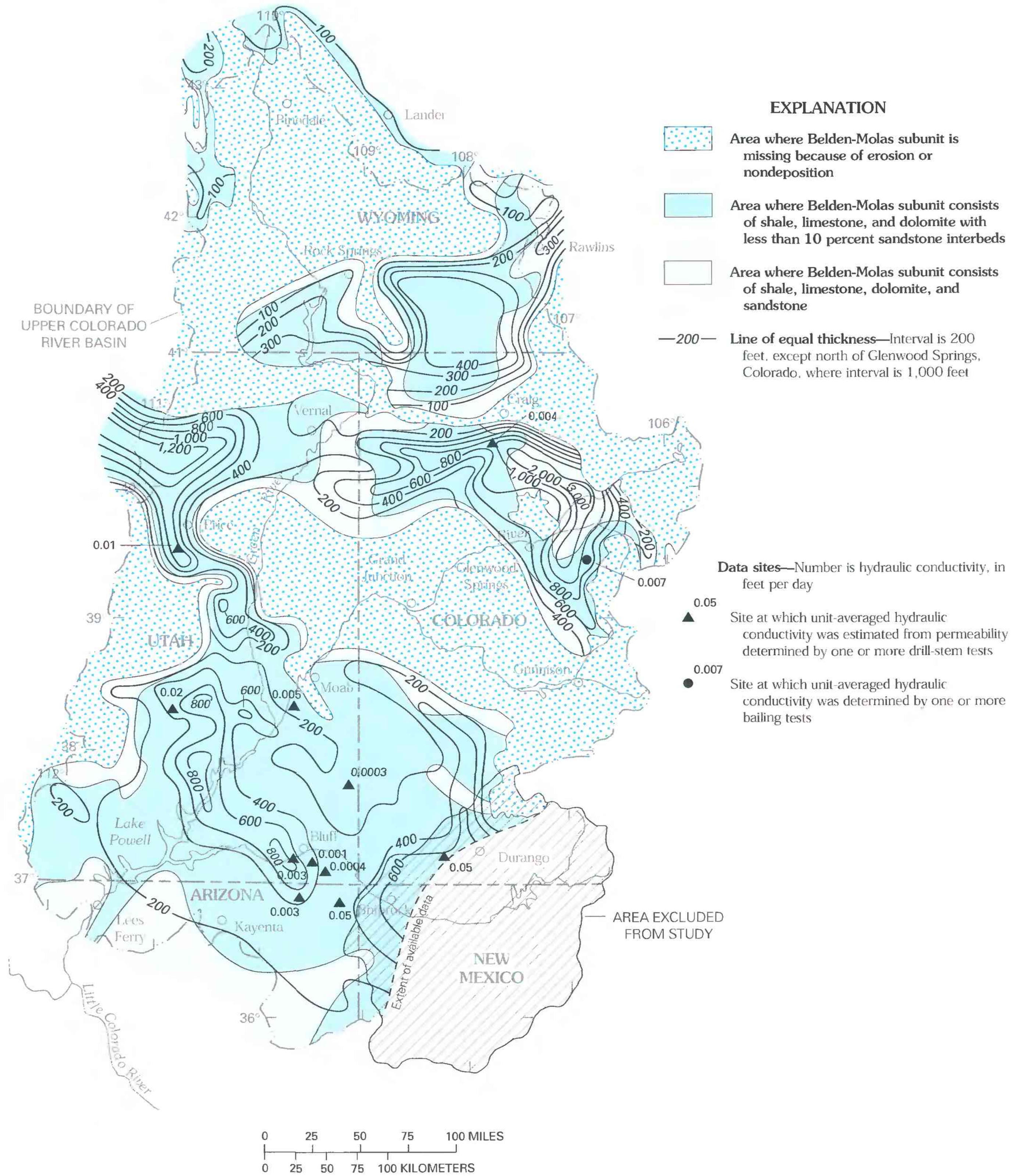


FIGURE 50.—Thickness, lithology, and unit-averaged hydraulic conductivity of the Belden-Molas subunit of the Four Corners confining unit. (Modified from Geldon, in press, pl. 13.)

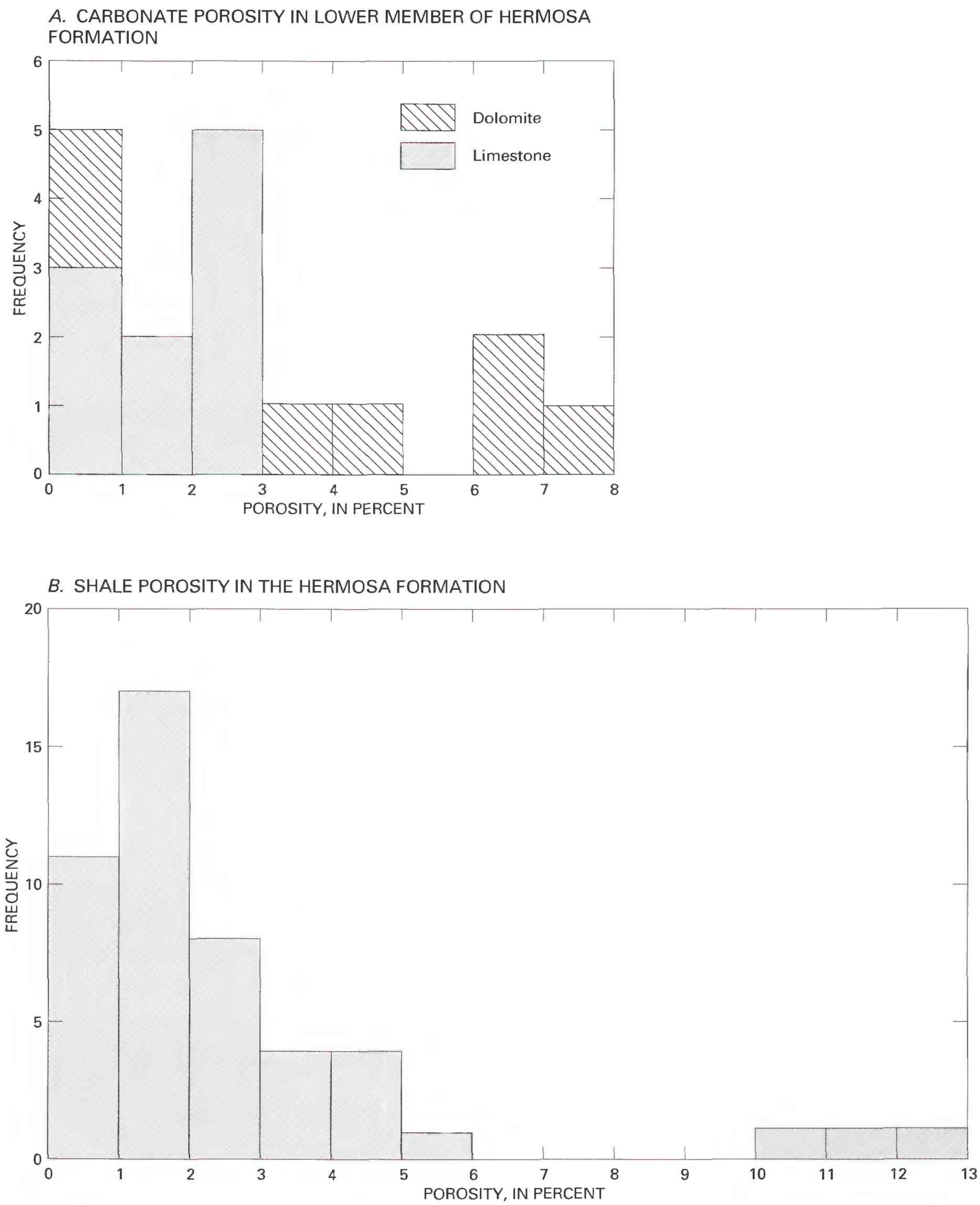


FIGURE 51.—Frequency distribution of porosity in carbonate rocks and shale from the Hermosa Formation.

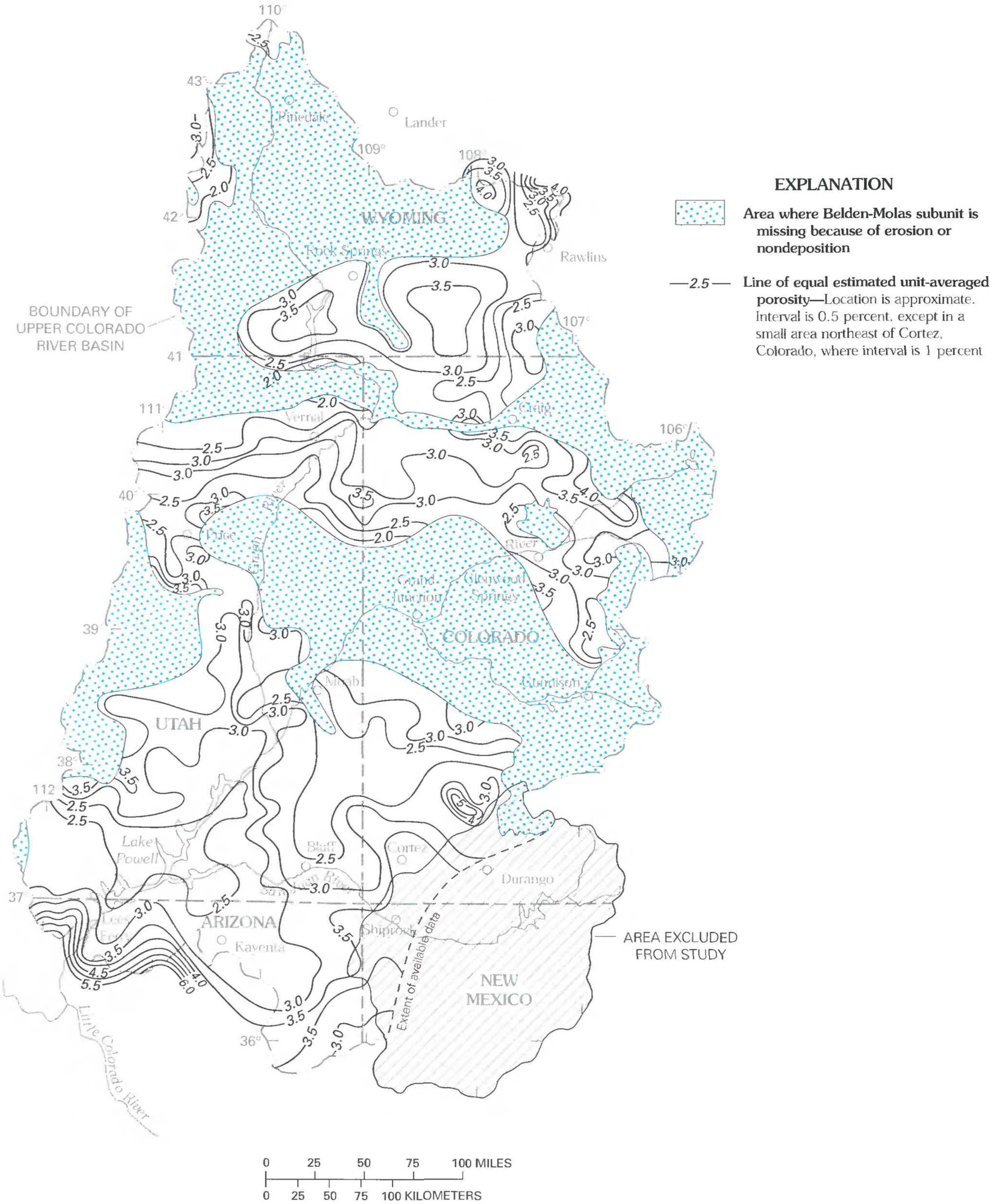


FIGURE 52.—Estimated distribution of unit-averaged porosity in the Belden-Molas subunit of the Four Corners confining unit.

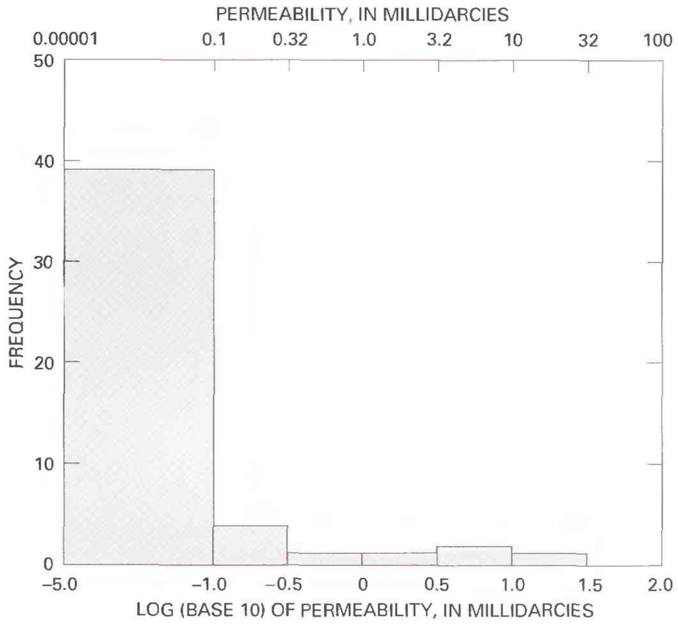


FIGURE 53.—Frequency distribution of pore-scale permeability in samples of shale from the Hermosa Formation.

ranged from 0.0012 to 130 md, with a median value of 1.1 md. Most of the tested intervals consisted of limestone or dolomite and shale; two intervals also contained sandstone.

HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

The drill-stem test data for the Belden-Molas subunit discussed in the previous section are equivalent to hydraulic-conductivity values with a range of 0.000029 to 0.32 ft/d. Bailing and pressure-injection tests at the Ruedi Dam site that were conducted by the Bureau of Reclamation (written commun., 1985) indicate additional values of hydraulic conductivity. Two pressure-injection tests in a 44-ft-thick interval of gypsiferous mudstone and siltstone of the Belden Formation in well DH–61 (SC08–84–17bac$_2$) indicated an average hydraulic-conductivity value of 0.54 ft/d. Nine pressure-injection tests in a 102-ft-thick interval of sandstone, siltstone, and mudstone of the Belden Formation in well DH–62 (SC08–84–17bac$_1$) indicated hydraulic-conductivity values ranging from 0.089 to 5.9 ft/d; a bailing test in this same well that was analyzed by the method of Skibitzke (1958) indicated a hydraulic-conductivity value of 0.073 ft/d. Residual drawdown data for a bailing test in well DH–13 (SC08–84–07caa), perhaps, indicate the most representative hydraulic-conductivity value for the Belden Formation in the Ruedi Dam area. These data, shown in figure 54, were obtained from an 86.4-ft-thick section of gypsiferous siltstone and claystone. During the test, 3.7 gallons were bailed in 225 minutes, which is equivalent to an average discharge of 0.016 gal/min, or 3.2 ft/d. The transmissivity calculated using equation 15 was:

$$T = \frac{35.2 \times 0.016 \text{ gal/min}}{(2.78 - 1.82)\text{ft}} = 0.59 \text{ ft}^2\text{/d},$$

and the hydraulic conductivity, calculated by dividing the transmissivity by the test-interval thickness, was 0.007 ft/d.

The complete range in hydraulic-conductivity values for the Belden-Molas subunit determined from 15 drill-stem tests, 3 laboratory determinations of permeability, the average of 2 injection tests in 1 borehole interval, and 2 bailing tests is 0.000013 to 0.54 ft/d, (fig. 55); the median of these hydraulic-conductivity values is 0.0046 ft/d. Keeping in mind that the data are skewed toward the relatively few aquifer tests in the Belden-Molas subunit that yielded water, the range in and median values for this hydrogeologic unit probably are smaller than what can be ascertained from the available data. In areas where the Belden-Molas subunit is mostly shale and deeply buried, the

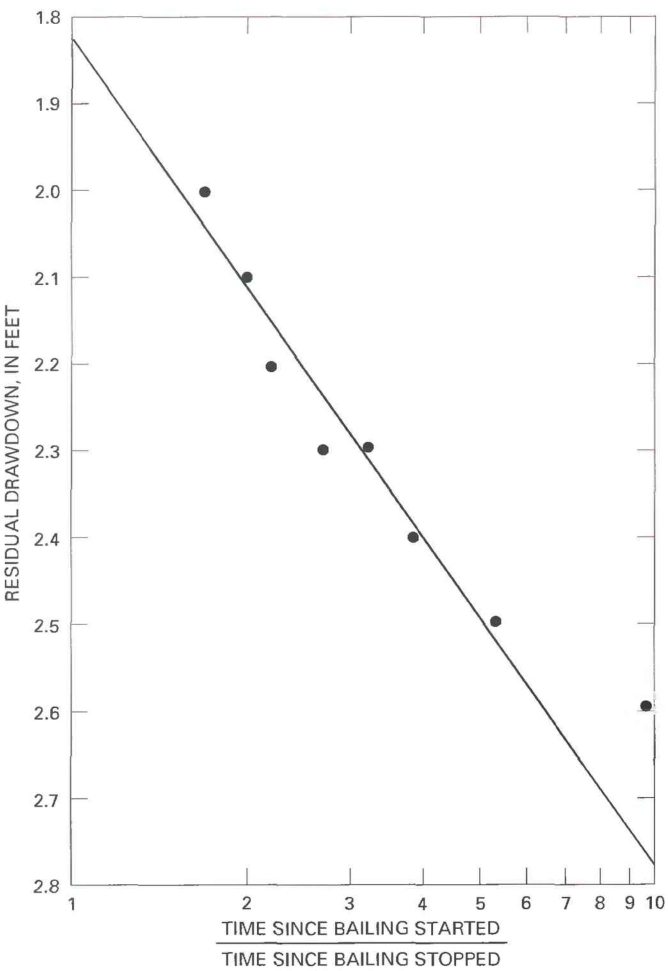


FIGURE 54.—Residual drawdown in well DH–13 (SC08–84–07caa) at Ruedi Dam site, Colorado, during bailing test of Belden Formation in March 1963 (data from Bureau of Reclamation, written commun., 1985).

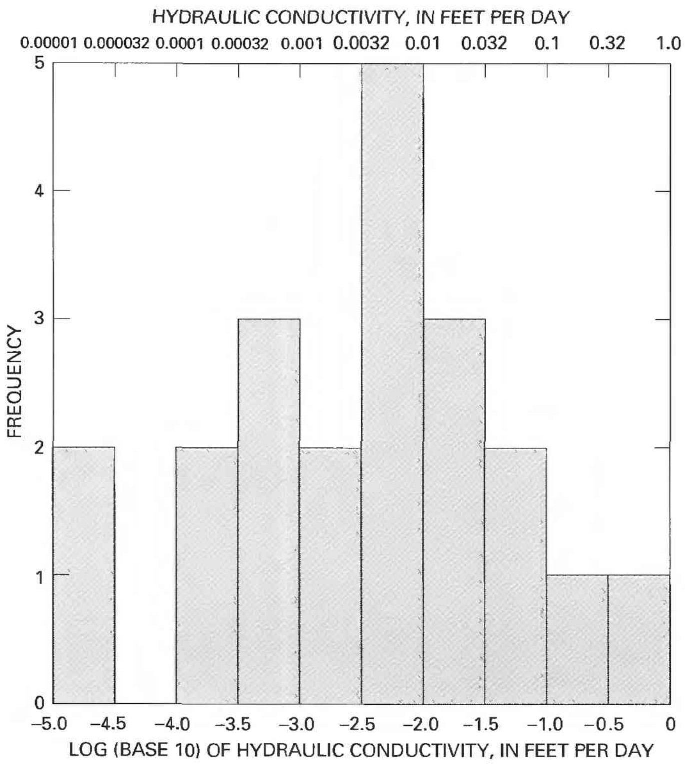


FIGURE 55.—Frequency distribution of hydraulic conductivity in the Belden-Molas subunit of the Four Corners confining unit.

hydraulic conductivity probably is negligible. Conversely, intervals of loosely consolidated sandstone and siltstone that are present in uplifted areas and shallow basins, such as the Eagle Basin, eastern Paradox Basin, and Kaibab Plateau, may have hydraulic-conductivity values that are larger than the maximum known value for the Belden-Molas subunit. On the basis of the limited available data, unit-averaged hydraulic conductivity is estimated to range from 0.00001 to less than 1.0 ft/d (fig. 55). The composite transmissivity of the Belden-Molas subunit, as determined from regional distributions of unit-averaged hydraulic conductivity and thickness and the bailing tests in wells DH–13 and DH–62, is estimated to range from 0.001 to 50 ft^2/d.

YIELDS FROM WELLS AND SPRINGS

Reflecting the generally small transmissivity, well and spring discharges from the Belden-Molas subunit rarely exceed 30 gal/min (fig. 56). In 15 drill-stem tests, yields from the Belden Formation, Doughnut Shale, Molas Formation, and lower member of the Hermosa Formation ranged from 1.0 to 30 gal/min, with a median value of 9.4 gal/min. Many other drill-stem tests, however, produced too little water to obtain a measurable discharge. During drilling of an exploration hole (SC08–84–14baa) at the Ruedi Dam site, water flowed from faulted and brecciated shale and limestone in the Belden Formation at a rate of more than 100 gal/min, but this discharge decreased to 8 gal/min by the time the hole was completed (Bureau of Reclamation, Denver, Colo., written commun., 1985). Four springs issuing from the Belden Formation at Dotsero, Colo., discharge at rates of 9 to 18 gal/min, but some of the water discharging from these springs may originate in the Leadville Limestone of the underlying Madison aquifer. The largest known yields from the Belden-Molas subunit in the UCRB are to two springs on the White River Plateau. These springs, according to Teller and Welder (1983), have moderate yields of 102 to 266 gal/min.

PARADOX–EAGLE VALLEY SUBUNIT

The Paradox–Eagle Valley subunit of the Four Corners confining unit consists of the Moffat Trail Limestone Member of the Amsden Formation, the Round Valley Limestone, the Eagle Valley Evaporite, the Paradox Member of the Hermosa Formation, and the Manakacha Formation of the Supai Group (table 1). Component geologic units are Late Mississippian to Middle Pennsylvanian in age. Together with the Belden-Molas subunit, the Paradox–Eagle Valley subunit so thoroughly retards ground-water movement that aquifers above and below are effectively isolated, except in areas of structural disturbance. In the Paradox Basin, for example, studies by Thackston and others (1981), Woodward-Clyde Consultants, Inc. (1982), and INTERA Environmental Consultants, Inc. (1984), concluded that water in the Paradox Member is isolated from any regionally connected flow system and that significant vertical and horizontal movement of water through this geologic unit is possible only where bedded evaporites that form most of the Paradox Member are penetrated by open fractures and faults or thinned by plastic deformation. Studies by the U.S. Geological Survey in the Paradox Basin (Rush and others, 1982; Weir and others, 1983a, 1983b; Whitfield and others, 1983) found that only 24 percent of all drill-stem tests in the Paradox Member produced enough water to calculate a meaningful discharge and that the Paradox Member was less permeable than all other geologic units of Cambrian to Tertiary age in the area. The Eagle Valley Evaporite, which is very similar to the Paradox Member, is presumed to be nearly as resistant to ground-water movement, although it locally supplies small quantities of fresh to brackish water to wells and springs (Brogden and Giles, 1976a; Giles and Brogden, 1976). Nothing is known about the water-bearing properties of the Round Valley Limestone, Moffat Trail Limestone Member, or Manakacha Formation, but an unspecified interval of cherty limestone in the Amsden Formation in the Overthrust Belt (if not the Moffat Trail Limestone Member, then a lithologically similar interval) discharges a small quantity of water to a well (Lines and Glass, 1975), and small springs issue from unspecified intervals in the Supai Group in the Grand Canyon (Metzger, 1961). Because the most important characteristic of this hydrogeologic unit is its

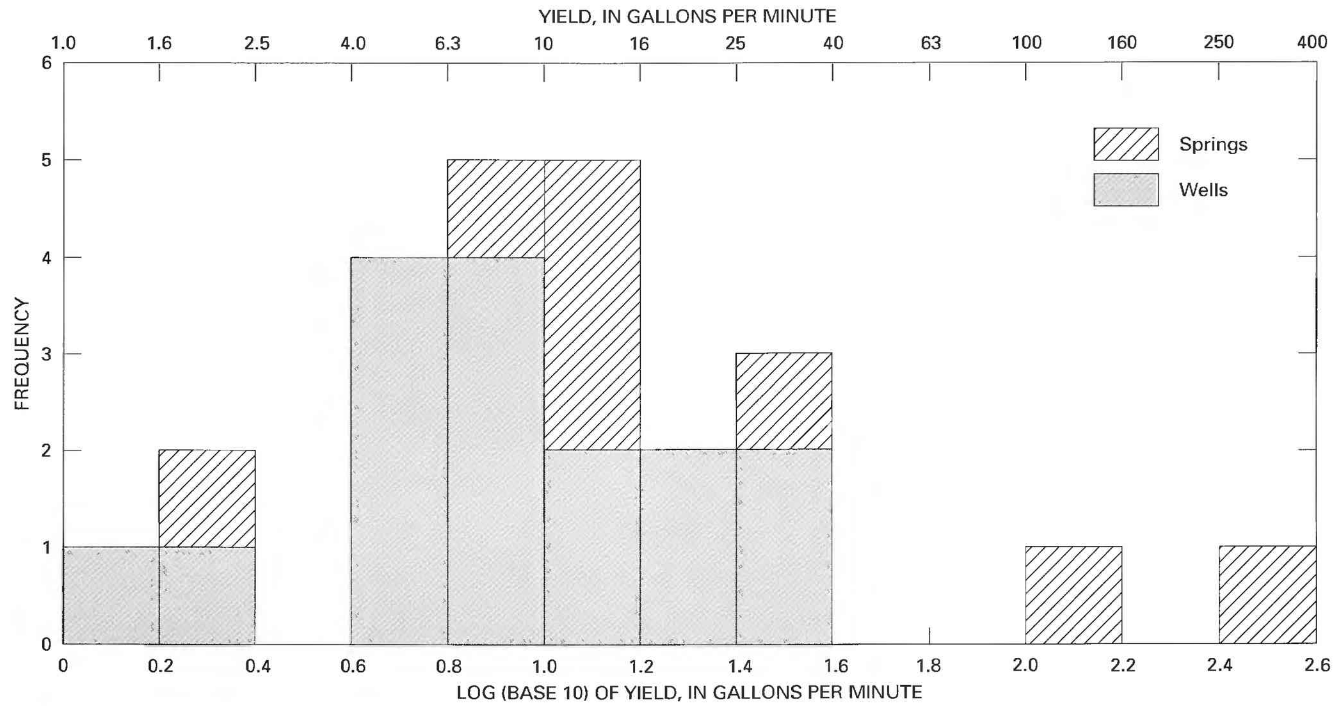


FIGURE 56.—Frequency distribution of yields from the Belden-Molas subunit of the Four Corners confining unit.

resistance to ground-water movement, the Paradox–Eagle Valley subunit was classified as part of a confining unit in this investigation.

THICKNESS AND LITHOLOGY

The thickness of the Paradox–Eagle Valley subunit ranges from 0 to more than 6,000 ft (fig. 57), except in small areas of the Paradox Basin where salt diapirs as much as 15,000 ft thick are present. Regionally, the Paradox–Eagle Valley subunit consists of varying proportions of limestone, dolomite, shale, sandstone, gypsum-anhydrite, and halite. Gypsum-anhydrite, halite, shale, and minor carbonate rocks are the dominant rock types in and near the Paradox and Eagle Basins. This hypersaline facies grades outward in and near the two basins into a penesaline facies composed of limestone, dolomite, shale, gypsum-anhydrite, and sandstone, which, in turn, grades outward into a marine-shelf facies composed of limestone and dolomite with subordinate shale and minor sandstone. The marine-shelf facies predominates from the Uinta and Sand Wash Basins to the northern edge of the UCRB. On the southern edge of the UCRB, a facies consisting of red shale and sandstone, subordinate carbonate rocks, and minor gypsum-anhydrite is present. Contacts between component geologic units and geologic units that comprise the overlying Canyonlands aquifer generally are gradational to disconformable.

POROSITY AND PERMEABILITY

Porosity in the Paradox–Eagle Valley subunit varies widely from place to place and within vertical sequences because of considerable variations in lithology. Sandstone and dolomite, with median porosities of about 6 to 7 percent, are the most porous rock types (table 9). On the basis of 176 analyses of samples from the Molas and Hermosa Formations (fig. 58), sandstone porosity in the Paradox–Eagle Valley subunit is estimated to range from 0.1 to 21 percent, with a median value of about 7 percent. In 729 analyses of dolomite from the Paradox–Eagle Valley subunit, porosity ranged from 0.1 to 31 percent, with a median value of 6.1 percent. In 1,649 analyses of limestone from the Paradox–Eagle Valley subunit, porosity ranged from less than 0.1 to 32 percent, with a median value of 3.5 percent. For both dolomite and limestone, median porosity in the Paradox–Eagle Valley subunit generally was found to be larger in varieties with a tendency toward dissolution around inclusions, such as cherty or vuggy limestone and cherty, anhydritic, or vuggy dolomite, than in shaly varieties (table 9). Median porosity also was larger in granular (sucrose, chalky, or earthy) varieties of limestone and dolomite than in fine-grained or crystalline varieties. The porosity of shale in the Paradox–Eagle Valley subunit is estimated from samples from the entire Hermosa Formation to range from less than 1 to 13 percent, with a median value of about 2 percent. The porosity of anhydrite and halite in the Paradox–Eagle Valley subunit, based on 19 analyses of samples from the Paradox Member, is estimated to range from 0.2 to 2.5 percent, with a

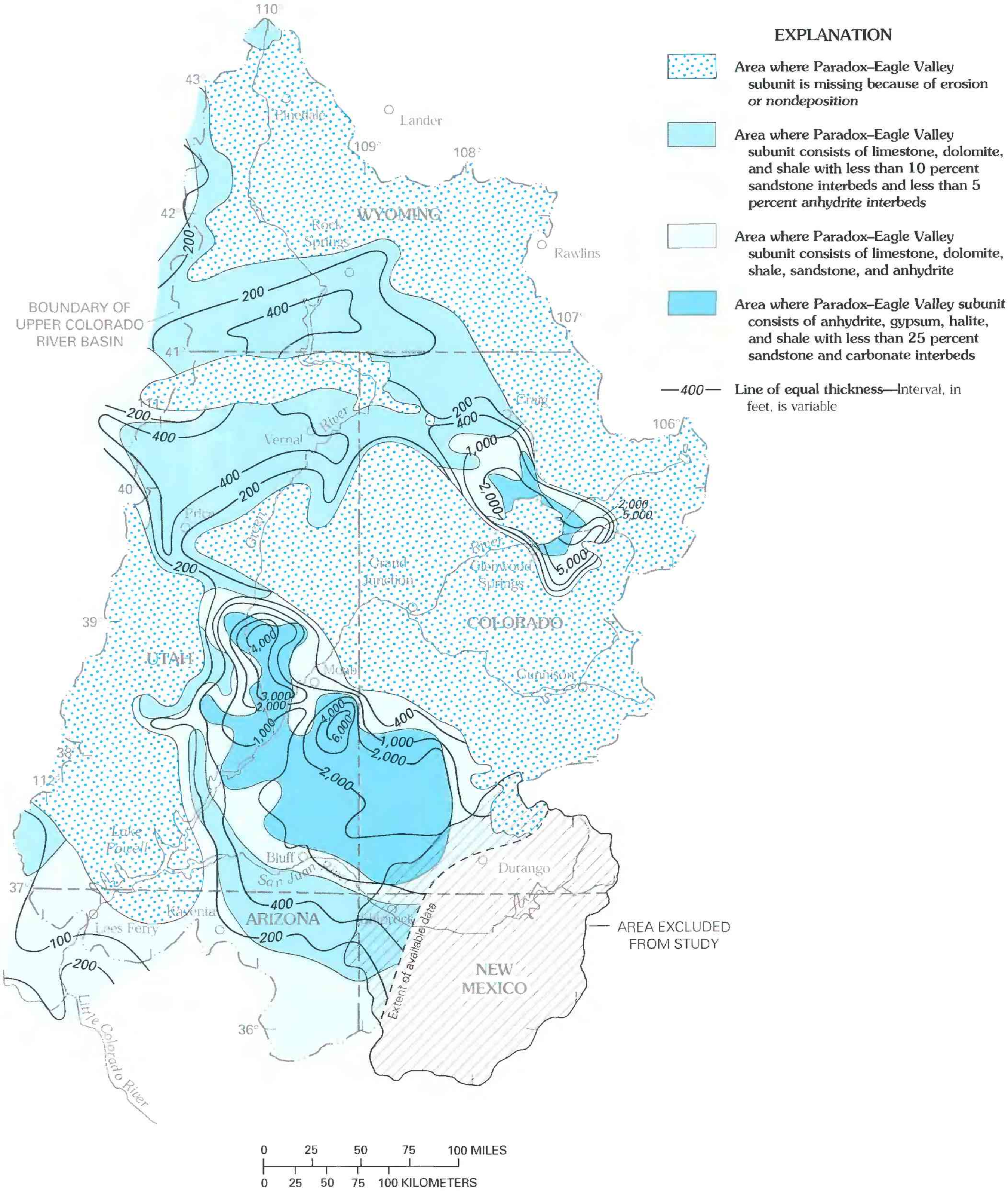


FIGURE 57.—Thickness and lithology of the Paradox–Eagle Valley subunit of the Four Corners confining unit. (Modified from Geldon, in press, pl. 14.)

TABLE 9.—*Porosity and pore-scale permeability statistics for the Paradox–Eagle Valley subunit of the Four Corners confining unit*

[<, less than]

Rock type	Porosity (percent) Minimum	Porosity (percent) Maximum	Porosity (percent) Median	Number of observations	Pore-scale permeability (millidarcies) Minimum	Pore-scale permeability (millidarcies) Maximum	Pore-scale permeability (millidarcies) Median	Number of observations
Halite	0.2	0.6	0.4	4	<0.0001	0.00022	0.0001	4
Anhydrite	.2	2.5	.4	15	.00017	27	.04	15
Shale[1]	.3	13	1.8	48	<.001	18	<.01	48
Sandstone[2]	.1	21	6.6	176	.00019	25	.01	176
Limestone								
Fossiliferous	.2	13.3	1.7	52	<.01	142	.05	52
Shaly	.3	19.0	2.5	158	<.01	142	.05	158
Anhydritic	.2	21.7	2.8	142	<.01	43	.17	142
Algal	.8	17.6	5.8	35	<.01	22	.70	35
Dolomitic	.3	27.3	6.3	177	<.01	84	.18	177
Vuggy	.3	28.5	6.8	238	<.01	2,435	1.3	238
Cherty	1.4	21.9	10.7	78	<.01	585	2.8	78
Brecciated	.5	4.3	2.4	9	.03	114	.20	9
Crystalline	<.1	18.1	2.5	618	<.01	2,435	.04	618
Fine grained	.3	27.3	3.0	92	<.01	54	.03	92
Sucrosic	.1	26.5	6.9	187	<.01	1,291	.79	187
Chalky	.4	16.8	10.1	9	<.01	6.9	.30	9
Earthy	2.4	19.7	12.0	12	.00039	4.4	.32	12
All	<.1	32.1	3.5	1,649	.00039	2,435	.10	1,649
Dolomite								
Shaly	.4	20.3	3.8	76	<.01	33	.01	76
Anhydritic	.6	22.6	10.2	86	<.01	1,644	4.9	86
Limy	.1	21.6	12.3	103	<.01	3,460	5.8	103
Cherty	2.0	24.6	12.6	21	<.01	699	.81	21
Vuggy	2.9	31.1	13.3	122	.04	3,460	31	122
Fine grained	.3	13.7	3.0	29	<.01	19	.04	29
Crystalline	.3	31.1	7.2	301	<.01	3,460	.59	301
Earthy	1.6	13.0	7.7	17	.01	1.5	.08	17
Sucrosic	.3	31.0	11.8	72	<.01	145	1.2	72
All	.1	31.1	6.1	729	<.01	3,460	.21	729

[1]Includes samples from all members of the Hermosa Formation.
[2]Includes samples from all members of the Hermosa Formation and the Molas Formation.

median value for each rock type of 0.4 percent. On the basis of lithologic composition and the above-cited median porosity values, unit-averaged porosity in the Paradox–Eagle Valley subunit is estimated to range regionally from less than 1 to more than 5 percent (fig. 59).

The permeability of the Paradox–Eagle Valley subunit varies from small to large, depending on the types of rocks present and the degree of fracturing and solution from area to area. Pore-scale permeability, which ranges from less than 0.0001 to 3,460 md, is largest in dolomite and limestone, intermediate in sandstone and anhydrite, and smallest in shale and halite (table 9). Carbonate textures affect pore-scale permeability in much the same way that they affect porosity. There appears to be no relation between pore-scale permeability and porosity for most rock types in this hydrogeologic unit (table 3). However, pore-scale permeability is related very crudely to porosity in limestone (fig. 60) and sandstone (table 3).

Reflecting the influence of secondary openings, median values of local-scale permeability are larger than median values of pore-scale permeability in carbonate, evaporite, and shale

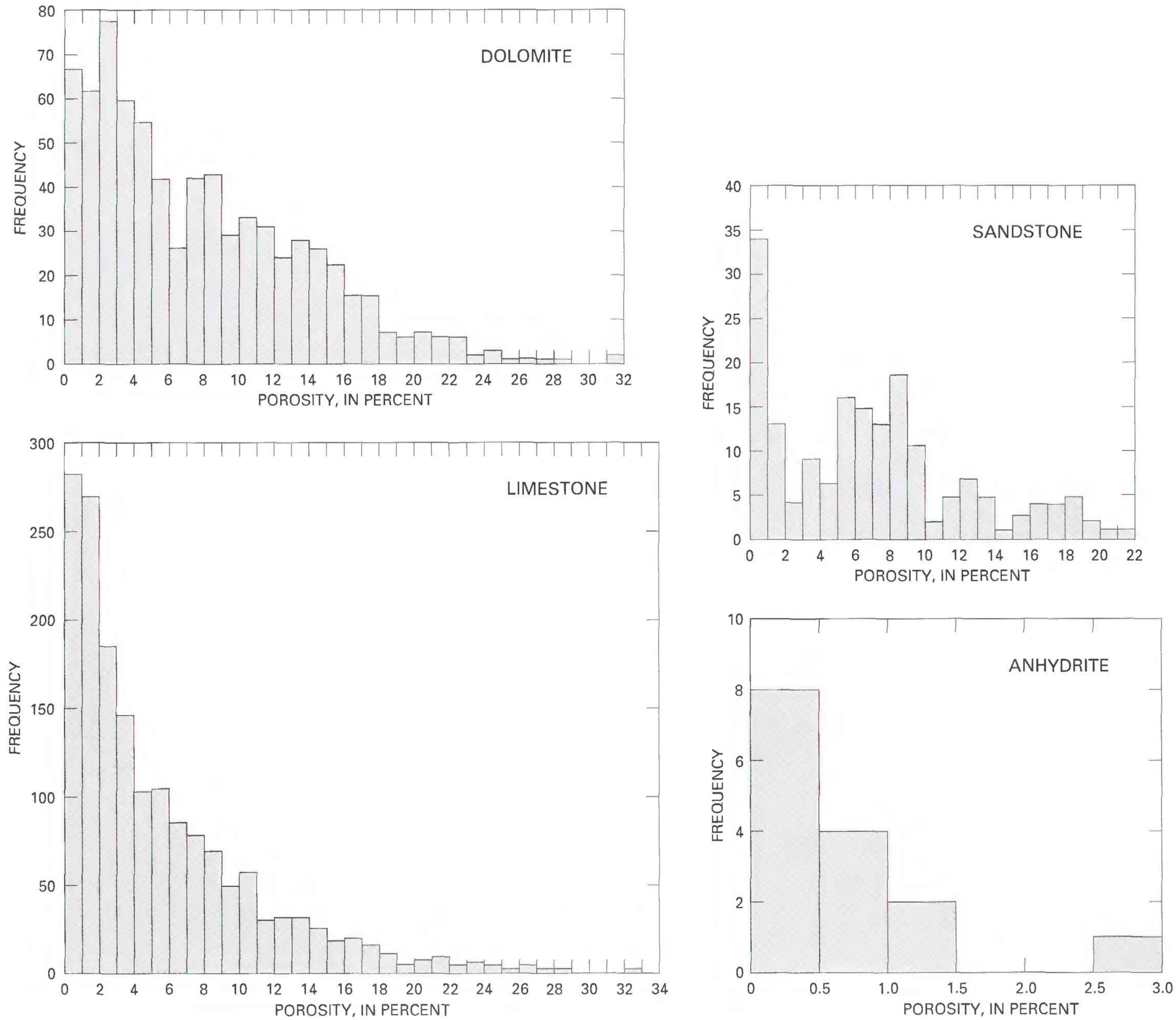


FIGURE 58.—Frequency distribution of porosity in carbonate rocks, sandstone, and anhydrite from the Paradox–Eagle Valley subunit of the Four Corners confining unit (the porosity of shale in this subunit can be inferred from fig. 51).

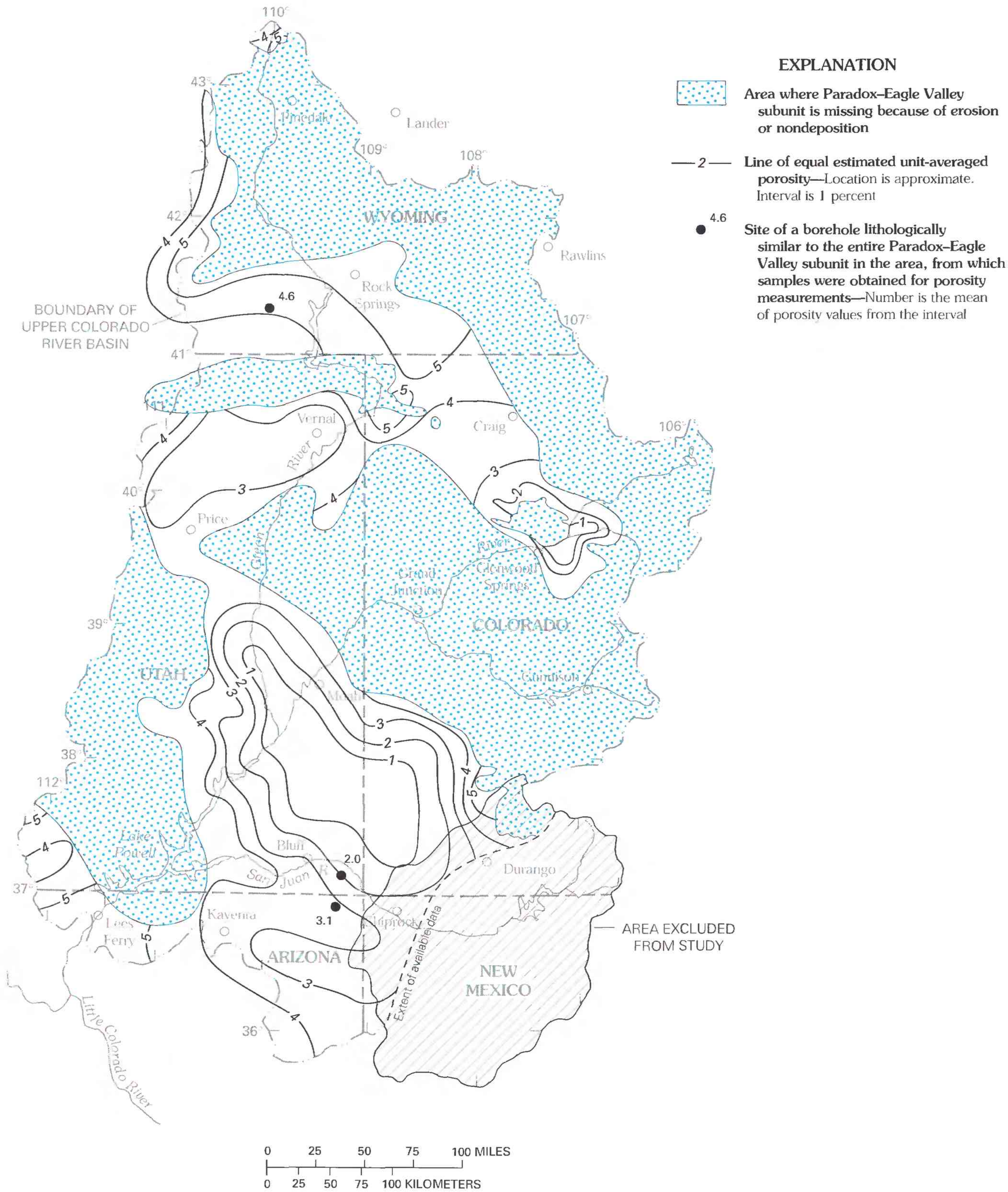


FIGURE 59.—Estimated distribution of unit-averaged porosity in the Paradox–Eagle Valley subunit of the Four Corners confining unit.

intervals. In 131 determinations (including 93 drill-stem tests, 3 slug tests, and 35 calculations from the average pore-scale permeability in a borehole interval), local-scale permeability ranged from 0.0031 to 550 md, with a median value of 2.3 md (fig. 61). In tested intervals consisting entirely of limestone and dolomite, local-scale permeability ranged from 0.017 to 350 md, with a median value of 7.1 md. In intervals consisting of carbonate rocks with shale, anhydrite, or halite interbeds, local-scale permeability ranged from 0.10 to 550 md, with a median value of 1.2 md. In intervals consisting entirely of shale, anhydrite, and halite, local-scale permeability ranged from 0.0031 to 33 md, with a median value of 1.2 md. The local-scale permeability of halite was found in six tests to range from less than 0.01 to 8.4 md, with a median value of 0.36 md.

HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

Values of hydraulic conductivity determined from the previously discussed permeability data and three slug tests ranged from 0.000008 to 1.5 ft/d (fig. 62), with median values of 0.018 ft/d for limestone and dolomite intervals, 0.0036 ft/d for carbonate intervals with shale and evaporite interbeds, and 0.0016 ft/d for intervals consisting of halite or anhydrite. On the basis of measurements of pore-scale permeability in a borehole in the Green River Basin (SB14–112–06bab), the unit-averaged hydraulic conductivity of the Round Valley Limestone and the Moffat Trail Limestone Member of the Amsden Formation in most areas can be expected to be near the median hydraulic conductivity of carbonate rocks in the Paradox–Eagle Valley subunit and range from 0.01 to 0.03 ft/d (pl. 4). Three drill-stem tests in the Paradox Basin confirm that predominantly carbonate parts of the Paradox Member of the Hermosa Formation and the Eagle Valley Evaporite also can be expected to have unit-averaged hydraulic-conductivity values between 0.01 and 0.03 ft/d. A drill-stem test near Meeker, Colo., and two drill-stem tests in the Henry Mountains Basin indicate that unit-averaged hydraulic-conductivity values of the Paradox Member and the Eagle Valley Evaporite should range from 0.003 to 0.007 ft/d where these geologic units consist of interbedded limestone, dolomite, sandstone, shale, and anhydrite. Drill-stem tests at six sites in the Paradox Basin indicate that unit-averaged hydraulic-conductivity values of the Paradox Member and the Eagle Valley Evaporite should range from 0.002 to 0.006 ft/d where these geologic units consist of interbedded carbonate rocks, anhydrite, and shale. Drill-stem tests at five sites in the Paradox Basin, including GD–1 (table 10), indicate that unit-averaged hydraulic-conductivity values of the Paradox Member and the Eagle Valley Evaporite should range from 0.001 to 0.005 ft/d where these geologic units consist mostly of interbedded anhydrite, gypsum, halite, and shale. Where the evaporites and interbeds are severely disturbed by diapiric intrusion, slug-injection tests at Salt Valley, in the Paradox Basin (table 11), indicate that unit-averaged hydraulic-conductivity values can be expected to range from less than 0.00001 to 0.00033 ft/d. Extrapolating from the above data,

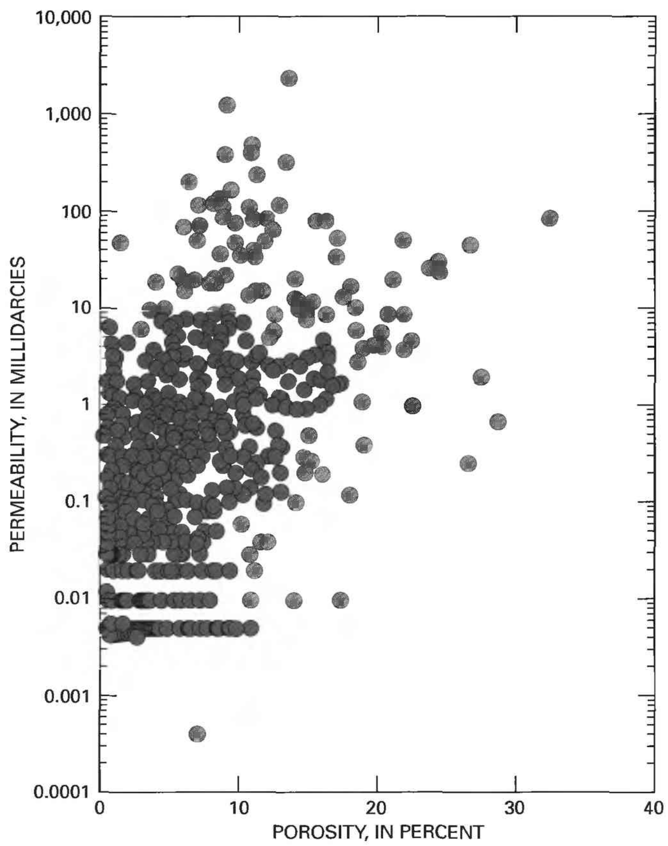


FIGURE 60.—Relation of porosity to pore-scale permeability in limestone samples from the Paradox–Eagle Valley subunit of the Four Corners confining unit.

unit-averaged hydraulic conductivity in the Paradox–Eagle Valley subunit, then, can be expected to range regionally from less than 0.0001 to about 0.03 ft/d (pl. 4).

Based mainly on regional distributions of thickness and unit-averaged hydraulic conductivity, the composite transmissivity of the Paradox–Eagle Valley confining unit is estimated to range from about 0.05 to 15 ft^2/d (pl. 4). These values are small to moderate. Because structural setting mostly was ignored in estimating composite transmissivity, actual values could be larger within and adjacent to uplifted areas, where all rock types tend to be fractured, and carbonate rocks tend to be cavernous. Actual values could be smaller in the center of structural basins because the effects of compaction were not considered. Given these limitations, the distribution of composite transmissivity appears to differ markedly north and south of the Uncompahgre Plateau. North of the plateau, there appears to be a progressive increase in transmissivity westward from the Eagle Basin and White River Plateau to the vicinity of the Uinta Mountains and Uinta Basin. This trend is caused by a lithologic change from predominantly shale and evaporites to predominantly carbonate rocks in the direction of increasing transmissivity. South of the Uncompahgre

Plateau, areas of alternately small and moderate transmissivity have a northwesterly alignment, in part because of lithofacies variations, but mainly because of the prevailing structural trend.

YIELDS FROM WELLS AND SPRINGS

Yields from the Paradox–Eagle Valley subunit to wells and springs typically are small, but moderate yields can occur. In 66 drill-stem tests, yields from the Paradox Member of the Hermosa Formation, Round Valley Limestone, and Eagle Valley Evaporite ranged from 0.21 to 103 gal/min (fig. 63). Many other drill-stem tests produced either too little water or too much mud, oil, or gas to calculate a water yield. According to Hood and Patterson (1984, p. 54), a petroleum test well in the San Rafael Swell flowed water at rates of 20 to 146 gal/min while being drilled. South of Glenwood Springs, Colo., wells developed in the Eagle Valley Evaporite yield water at rates of 5 to 25 gal/min with unspecified drawdowns (Brogden and Giles, 1976a). Springflows from this hydrogeologic unit occur rarely. Stinking Springs (SLD24–24–21c), which issues from the Paradox Member of the Hermosa Formation in the Paradox Basin, has a discharge of 20 to 30 gal/min (Ritzma and Doelling, 1969, p. 106); Onion Creek Spring (SLD24–24–22), in the same area, has a discharge of 55 gal/min (Rouse, 1967, p. 19). Tripp Hot Spring (NMB36–09–10bcb) and Trimble Hot Spring (NMB36–09–15bcb) near Durango, Colo., issue from the Paradox Member through overlying colluvium with

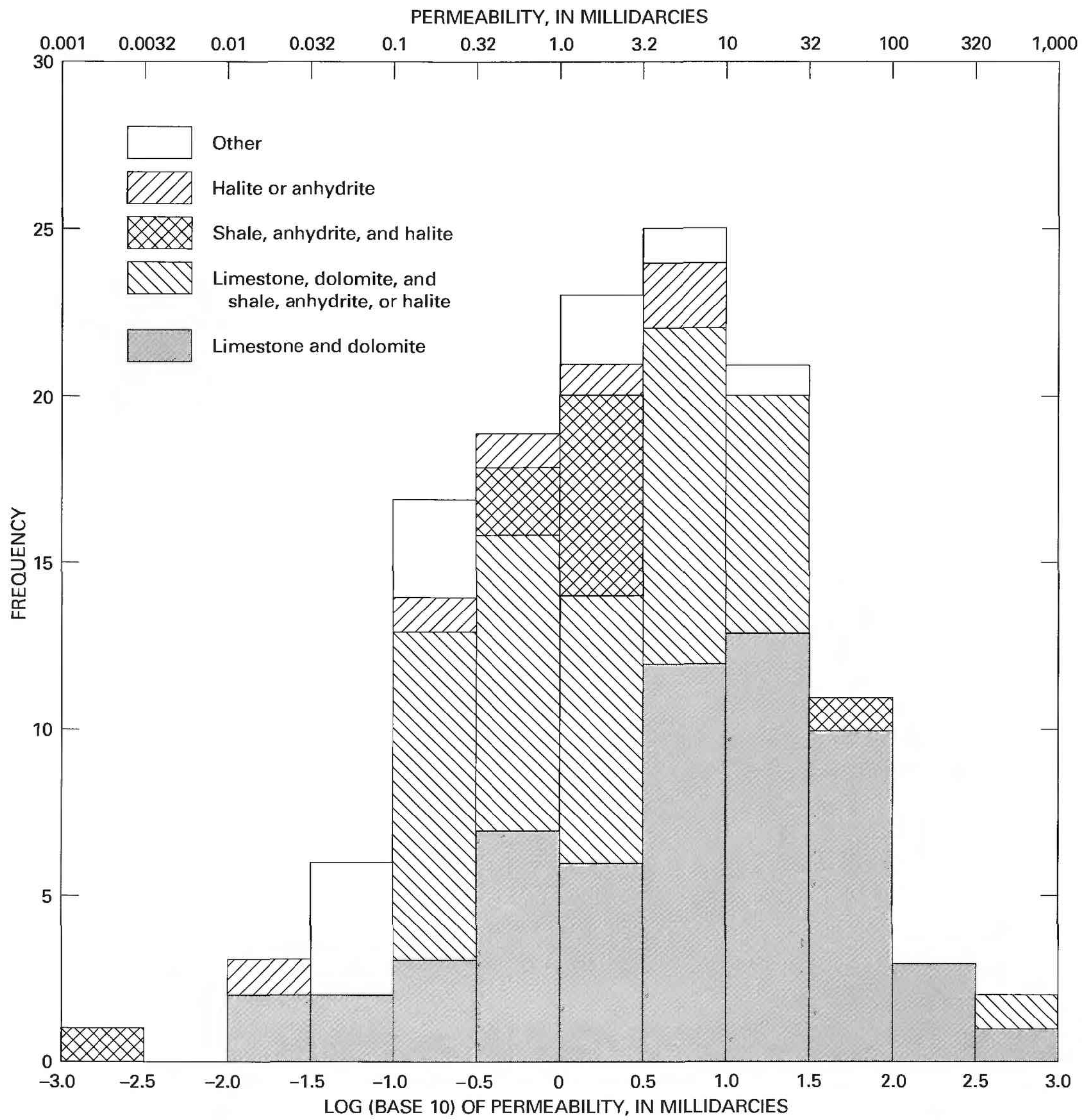


FIGURE 61.—Frequency distribution of local-scale permeability in the Paradox–Eagle Valley subunit of the Four Corners confining unit.

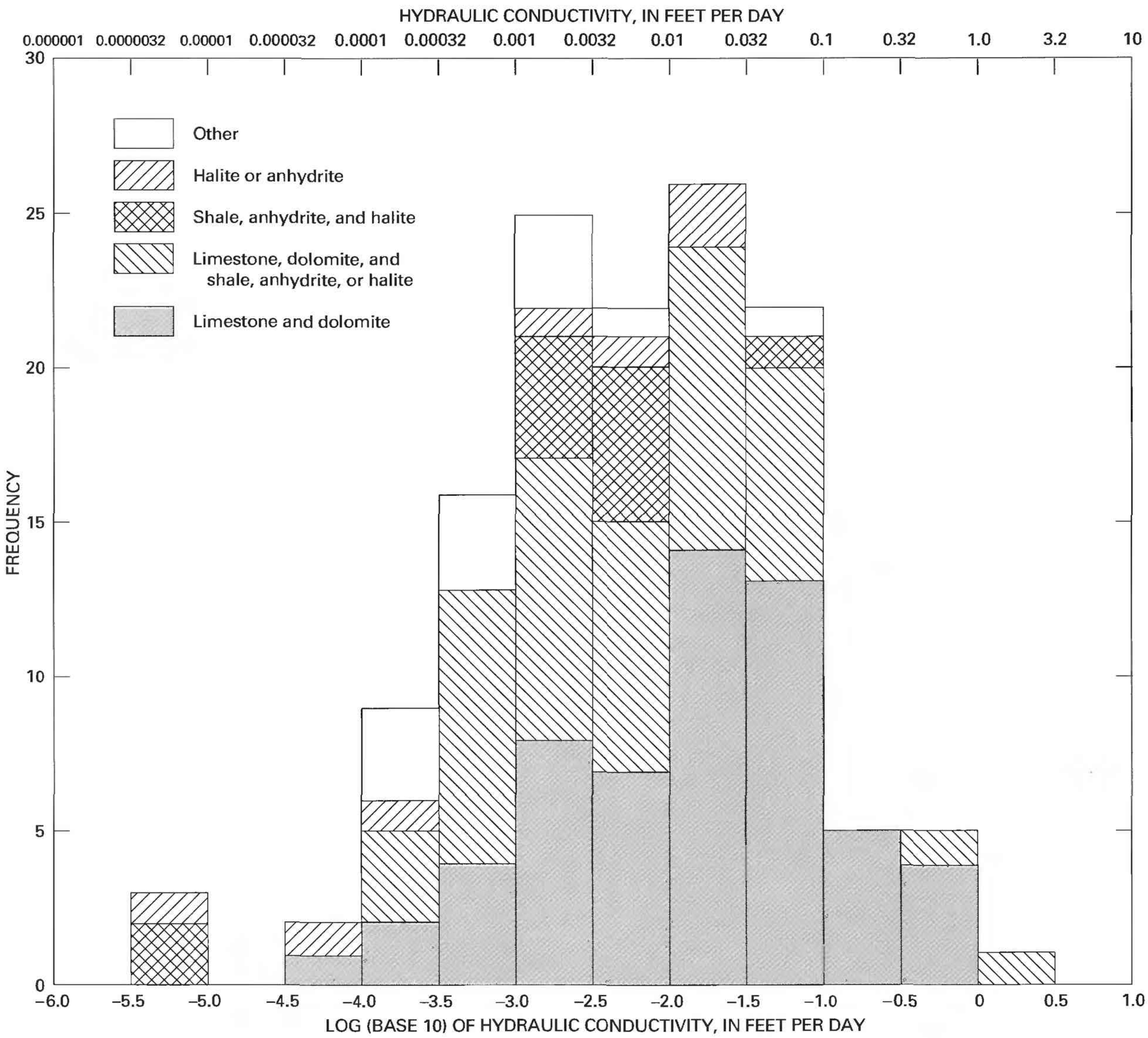


FIGURE 62.—Frequency distribution of hydraulic conductivity in the Paradox–Eagle Valley subunit of the Four Corners confining unit.

discharges of 15 to 24 gal/min (Barrett and Pearl, 1977, p. 240–242). Big Sulphur Spring (SC08–84–16aa), which issues from the Eagle Valley Evaporite near Meredith, Colo., has a discharge of 150 gal/min (Rouse, 1967, p. 14). Another spring issuing from the Eagle Valley Evaporite near Dotsero, Colo., Big Spring (SC05–86–05), was reported by Iorns and others (1964) to have a discharge of 450 gal/min. However, water issuing from this spring probably originates in the Leadville Limestone and rises to the surface through interconnecting faults. Under most circumstances, discharges to wells and springs from the Eagle Valley Evaporite and other component geologic units of the Paradox–Eagle Valley subunit are not likely to exceed 150 gal/min.

HYDROLOGIC PROPERTIES OF THE CANYONLANDS AQUIFER

Rocks equivalent to the C multiple-aquifer of the Navajo and Hopi Indian Reservations in Arizona, New Mexico, and Utah (Cooley and others, 1969) extend throughout the UCRB. The C multiple-aquifer extends from the upper part of the Supai Group to the Kaibab Formation and, thus, includes all Paleozoic rocks above the Four Corners confining unit. To be consistent with current guidelines of the U.S. Geological Survey for naming hydrogeologic units (Laney and Davidson, 1986), the C multiple-aquifer herein is named the Canyonlands aquifer

TABLE 10.—*Hydraulic conductivity of the Paradox Member of the Hermosa Formation in Borehole GD–1 (SLD30–21–21ddd)*

[Drill-stem test data were compiled from Woodward-Clyde Consultants, Inc., 1982, table 9–2. Location of borehole shown on plate 4]

Interval below Kelly bushing[1] (feet)	Lithology	Hydraulic conductivity[2] (foot squared per day)
2,540–2,600	Limestone	0.000046
2,600–2,760	Dolomite, anhydrite, shale	(.0004)
2,760–2,960	Limestone, dolomite, anhydrite, shale	.00042
2,960–3,100	Limestone, dolomite, anhydrite, shale	.00065
3,100–3,330	Anhydrite, shale, halite	(.004)
3,330–3,530	Anhydrite, shale, halite	.0041
3,530–3,660	Halite	(.002)
3,660–3,800	Anhydrite	.019
3,800–3,940	Anhydrite, shale, halite	.0051
3,940–4,030	Halite	(.002)
4,030–4,230	Anhydrite shale, halite	.0030
4,230–4,730	Halite with anhydrite and shale	(.002)
4,730–4,950	Anhydrite, shale, halite, limestone	(.003)
4,950–5,090	Anhydrite, shale, halite	.0011
5,090–5,200	Halite	(.002)
5,200–5,340	Anhydrite, limestone, shale	.013
5,340–5,450	Halite with minor interbeds	(.002)
	Estimated average	.0036

[1]Altitude of Kelly bushing is 4,949.2 feet above NGVD of 1929.

[2]Unless enclosed in parentheses, value was obtained from a drill-stem test. Values in parentheses are estimates based on the median hydraulic conductivity of the rock types in the interval or the measured value for an interval of similar lithology in the borehole.

because of its importance as a source of water in the Canyonlands region of Utah, as well as in other deeply dissected areas of the UCRB.

The Canyonlands aquifer vertically is divisible into three zones with different lithologic and hydrologic properties. The lowermost zone, the Cutler-Maroon zone, consists mostly of interbedded clastic and carbonate rocks and is named after the two geologic units within the zone most characteristic of its lithology and most often used as sources of water. The middle zone, the Weber-De Chelly zone, consists almost entirely of quartz sandstone and is named for the two most prominent water-bearing formations within this zone. The uppermost zone, the Park City–State Bridge zone, consists alternately of carbonate rocks, carbonate and clastic rocks, or shale and is named for the two geologic units most representative of the extensive regional variations in lithology within the uppermost zone.

CUTLER-MAROON ZONE

The Cutler-Maroon zone of the Canyonlands aquifer consists of a very thick sequence of variably permeable clastic and carbonate rocks of Early Pennsylvanian to Late Permian age that are hydraulically connected by fractures associated with folds, faults, and igneous intrusions. Component geologic units include the Ranchester Limestone Member of the Amsden Formation; the upper member of the Hermosa Formation; the Gothic, Minturn, Morgan, Rico, Supai, and Cutler Formations; the Elephant Canyon Formation, Halgaito Shale, Cedar Mesa Sandstone, and Organ Rock Shale of the Cutler Group; the Wescogame Formation and Esplanade Sandstone of the Supai Group; the main body of the Maroon Formation (excluding the Schoolhouse and Fryingpan Members); and the Hermit Shale (table 1).

The most permeable component geologic unit is the Cedar Mesa Sandstone, which has both intergranular and fracture permeability (Richter, 1980, p. 19). The lower 150 feet of this formation is the source of numerous small to moderately sized springs in the ruggedly dissected Canyonlands area of southeastern Utah (Ritzma and Doelling, 1969, p. 64–65; Sumsion and Bolke, 1972, p. 48; Richter, 1980, p. 9–11). The Cedar

TABLE 11.—*Hydraulic conductivity of the Paradox Member of the Hermosa Formation in Borehole DOE–1 (SLD23–20–05bad$_1$)*

[Slug test data were compiled from Rush and others, 1980. Location of borehole shown on plate 4; <, less than]

Interval below land surface (feet)	Lithology	Hydraulic conductivity (feet per day)
626–738	Salt	[1](0.0001)
738–771	Interbeds	[2](.0003)
771–918	Salt	[1](.0001)
918–951	Interbeds	.00033
951–1,063	Salt	.00013
1,063–1,076	Interbeds	.000033
1,076–1,187	Salt	
1,187–1,197	Interbeds	
1,197–1,276	Salt	
[3]1,276–4,276	Salt with interbeds	[4](<.00001)
	Estimated average	.00003

[1]Estimate based on data for interval 951–1,063 feet.

[2]Estimate based on data for interval 918–951 feet.

[3]Total thickness estimated from borehole DOE–3 (SLD23–20–05bad). This borehole penetrated 3,520 feet of the Paradox Member without reaching the base of this unit.

[4]Estimate based on a slug test in DOE–3 between depths of 728 and 4,061 feet. The tested interval consisted of salt with interbeds. No systematic decline in water level was detected, indicating a very small hydraulic conductivity.

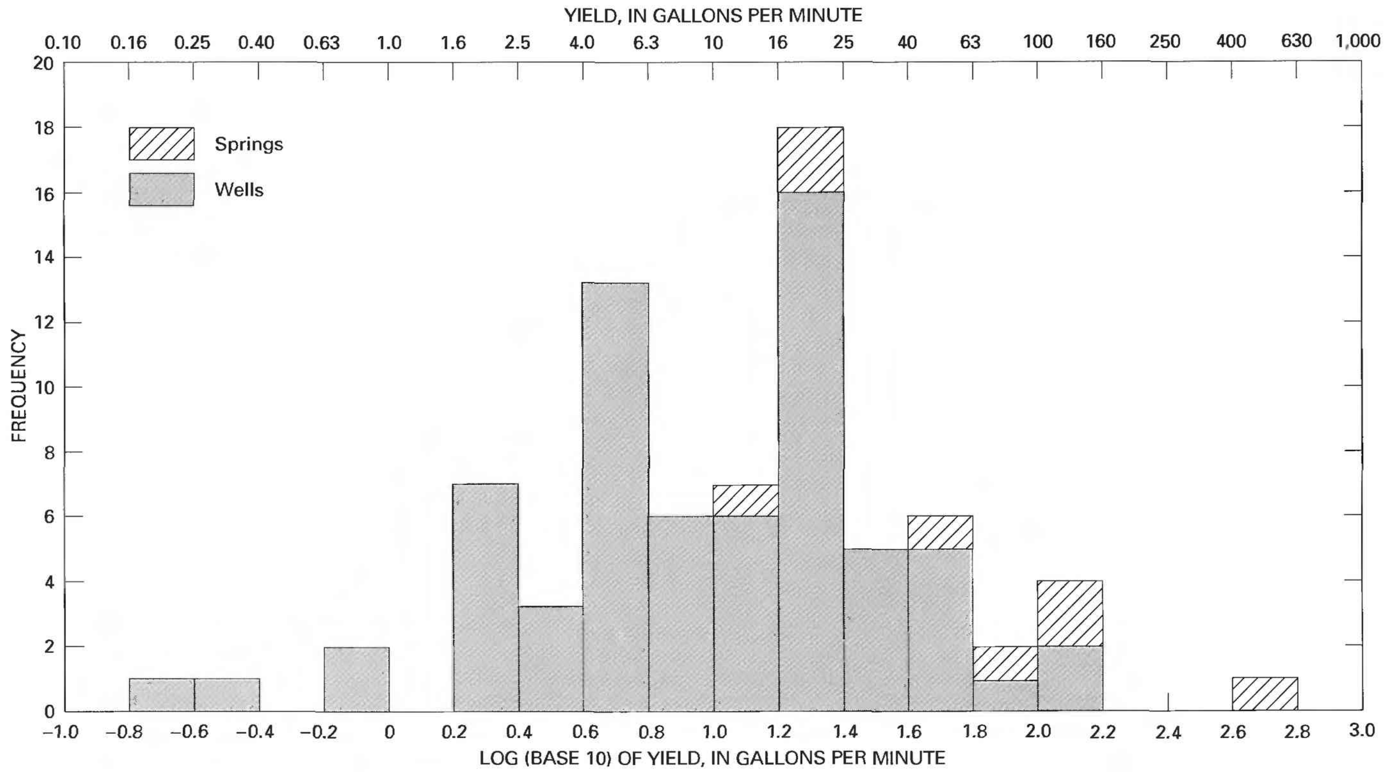


FIGURE 63.—Frequency distribution of yields from the Paradox–Eagle Valley subunit of the Four Corners confining unit.

Mesa Sandstone also yields small to moderate quantities of water to wells in the Canyonlands area, including supply wells in Canyonlands National Park (Sumsion and Bolke, 1972, p. 52) and Natural Bridges National Monument (Huntoon, 1977, p. 19–20; Hand, 1979, p. 7–8).

Component geologic units of the Cutler-Maroon zone that mostly consist of interbedded sandstone and shale are not as permeable as the Cedar Mesa Sandstone but supply water throughout a much larger area. Within these units, including the Minturn Formation, Gothic Formation, Cutler Formation, Supai Formation, main body of the Maroon Formation, and the Supai Group, generally small to moderate discharges of water to wells and springs occur where downward percolation is prevented by discontinuous layers of shale, fine-grained sandstone, or carbonate rocks. Most of the water transmitted by these geologic units is from intervals of sandstone or conglomerate, which have both intergranular and fracture permeability. Other rock types have mostly fracture or solution permeability (see Richter, 1980, p. 13). Of the geologic units comprising this category, the main body of the Maroon Formation is most widely used for water supplies (Brodgen and Giles, 1976a; Giles and Brogden, 1976; Wright Water Engineers (1979). The Maroon Formation is the principal bedrock aquifer supplying water to wells in the Roaring Fork Valley between Glenwood Springs and Aspen, Colo., and is the source of numerous springs in the Elk Mountains, including Arsenic Spring (SC11–87–35) and Conundrum Hot Springs (SC12–85–16ac). The Minturn Formation supplies small quantities of water to wells in the McCoy and Minturn areas of Colorado (Colorado Division of Water Resources, Office of the State Engineer, unpublished well permits). The Gothic Formation supplies small quantities of water to springs near Crested Butte, Colo. (Giles, 1980). The Cutler Formation supplies small to moderate quantities of water to wells and widely scattered springs on the edges of the San Juan Mountains and Uncompahgre Plateau and in the Paradox Basin (Sumsion, 1971, p. 37–39; Richter, 1980, p. 9, 20; Whitfield and others, 1983, p. 62–88; U.S. Geological Survey, unpublished data). The Supai Group yields small quantities of water to wells and springs in the Defiance Plateau area but generally does not transmit water readily (Akers and others, 1962, p. 3); in the Grand Canyon area, for example, only one small spring is known to issue from this unit (Metzger, 1961, p. 118).

Component geologic units of the Cutler-Maroon zone that consist of interbedded carbonate rocks, sandstone, and shale, including the Morgan, Elephant Canyon, and Rico Formations, the upper member of the Hermosa Formation, and the Ranchester Limestone Member of the Amsden Formation, have little intergranular permeability and transmit water mainly through fractures and solution channels (Richter, 1980, p. 13). Within these geologic units, sandstone and sandy limestone layers are the principal sources of water, particularly where they are underlain by shale or fine-grained limestone. Geologic units in this category are not often used to obtain water because water-bearing properties vary considerably and are unpredictable. The Morgan Formation, for example, generally is not water bearing, but it supplies small to large quantities of water to springs in the Jones Hole area at the eastern end of the Uinta Mountains (Sumsion, 1976, p. 45) where the formation is cut by faults and the overlying Weber-De Chelly zone is unsaturated. The Elephant Canyon and Rico Formations function mainly as a basal confining layer for the Cedar Mesa Sandstone; but in the Canyonlands area, small springs issue from fractured limestone layers in the upper 50 feet of these formations (Huntoon, 1979; Richter, 1980, p. 9). The upper member of the Hermosa Formation typically supplies small quantities of water to wells and springs throughout southeastern Utah and southwestern Colorado (Richter, 1980, p. 21–23; Thackston and others, 1981, p. 205), but large discharges can occur in fault zones where the permeability has been enhanced by fracturing (Huntoon, 1977, p. 7–8). According to Lines and Glass (1975), small discharges are possible from the Ranchester Limestone Member of the Amsden Formation, even though spring discharges are unknown, and few water wells are completed in this unit.

Component geologic units of the Cutler-Maroon zone that mostly consist of shale and very fine-grained sandstone, including the Organ Rock, Halgaito, and Hermit Shales, are negligibly permeable, even where transected by fractures. These units yield little or no water in most areas, functioning mainly as confining layers for the Cedar Mesa Sandstone and the Coconino, White Rim, and De Chelly Sandstones of the overlying Weber-De Chelly zone (Metzger, 1961, p. 118; Richter, 1980, p. 13; Rush and others, 1982, p. 15). However, a few small springs are known to issue from the Halgaito Shale in the Monument Upwarp (Ritzma and Doelling, 1969, p. 78–79).

The Cutler-Maroon zone has a tendency to produce water everywhere that it occurs, but because of thick intervals of negligibly permeable rock within this zone, wells hundreds to thousands of feet deep may produce little or no water. Because the depth at which water will be present in a well and the amount of discharge are unpredictable, the Cutler-Maroon zone can be expected to function subregionally as an aquifer in the UCRB.

THICKNESS AND LITHOLOGY

The thickness of the Cutler-Maroon zone ranges from 0 to more than 10,000 ft (fig. 64). Regionally, this zone consists of highly variable proportions of sandstone, conglomerate, shale, limestone, dolomite, and gypsum-anhydrite as a result of abrupt facies changes related to syndepositional uplift and subsidence. Sandstone and conglomerate generally make up 50 to more than 75 percent of the Cutler-Maroon zone within and adjacent to the Sawatch and Gore Ranges, the Elk and San Juan Mountains, the White River, Uncompahgre, Defiance, and Kaibab Plateaus, and the Paradox Basin. The sandstone in these areas predominantly is red, arkosic, micaceous, coarse grained, and friable; most of the interbeds are siltstone or mudstone; carbonate intervals are thin and sparsely distributed. Away from these areas, the rocks become predominantly gray, green, tan, and pink; arkosic sandstone layers become finer grained, more indurated, and less numerous; quartz sandstone intervals thicken and predominate over arkosic intervals; siltstone and mudstone intervals thicken and equal or exceed sandstone in abundance; limestone and dolomite intervals thicken and compose as much as 45 percent of the zone; thin gypsum or anhydrite beds occur locally. In northern and central parts of the UCRB, the Cutler-Maroon zone consists of limestone and dolomite with subordinate shale interbeds, generally less than 25 percent sandstone interbeds, and locally, minor interbeds of gypsum or anhydrite. Contacts between component geologic units and overlying Pennsylvanian and Permian rocks generally are conformable to gradational, but local unconformities exist. In southeastern Utah east of the Monument Upwarp and in western Colorado south of the Piceance Basin, component geologic units of the Cutler-Maroon zone generally are the youngest Paleozoic rocks and are overlain conformably to unconformably by Mesozoic rocks.

POROSITY AND PERMEABILITY

The porosity of the Cutler-Maroon zone varies considerably from place to place and within vertical sequences because of extreme variations in lithology. Sandstone is the most porous rock type within this zone. In 299 analyses, sandstone porosity ranged from 0.8 to about 20 percent, with a median value of 9.1 percent (fig. 65). The median porosity of sandstone varieties generally increases with decreasing cement and clay content and increasing grain size (table 12). Limestone and dolomite, on the average, are about one-third as porous as sandstone. In 210 analyses, the porosity of limestone and dolomite ranged from 0.4 to 16 percent, with median values of 2.4 percent for limestone and 3.3 percent for dolomite (fig. 65). In 10 analyses, shale porosity ranged from 0.5 to about 12 percent, with a median value of 4.7 percent, but given the small number of analyses, these figures probably are not statistically significant. On the basis of the porosity of shale in all of the hydrogeologic units composed of Paleozoic rocks, it is estimated that median shale porosity in the Cutler-Maroon zone probably is between 2 and 4 percent. On the basis of lithologic composition, median porosity values for common rock types, and sparsely distributed laboratory and geophysically determined site averages, unit-averaged porosity in the Cutler-Maroon zone is estimated to range regionally from about 2 to 14 percent (pl. 5).

Borehole geophysical logs show no obvious relation between porosity and depth but confirm a strong relation between porosity and lithology. In figure 66*A*, a geophysical log

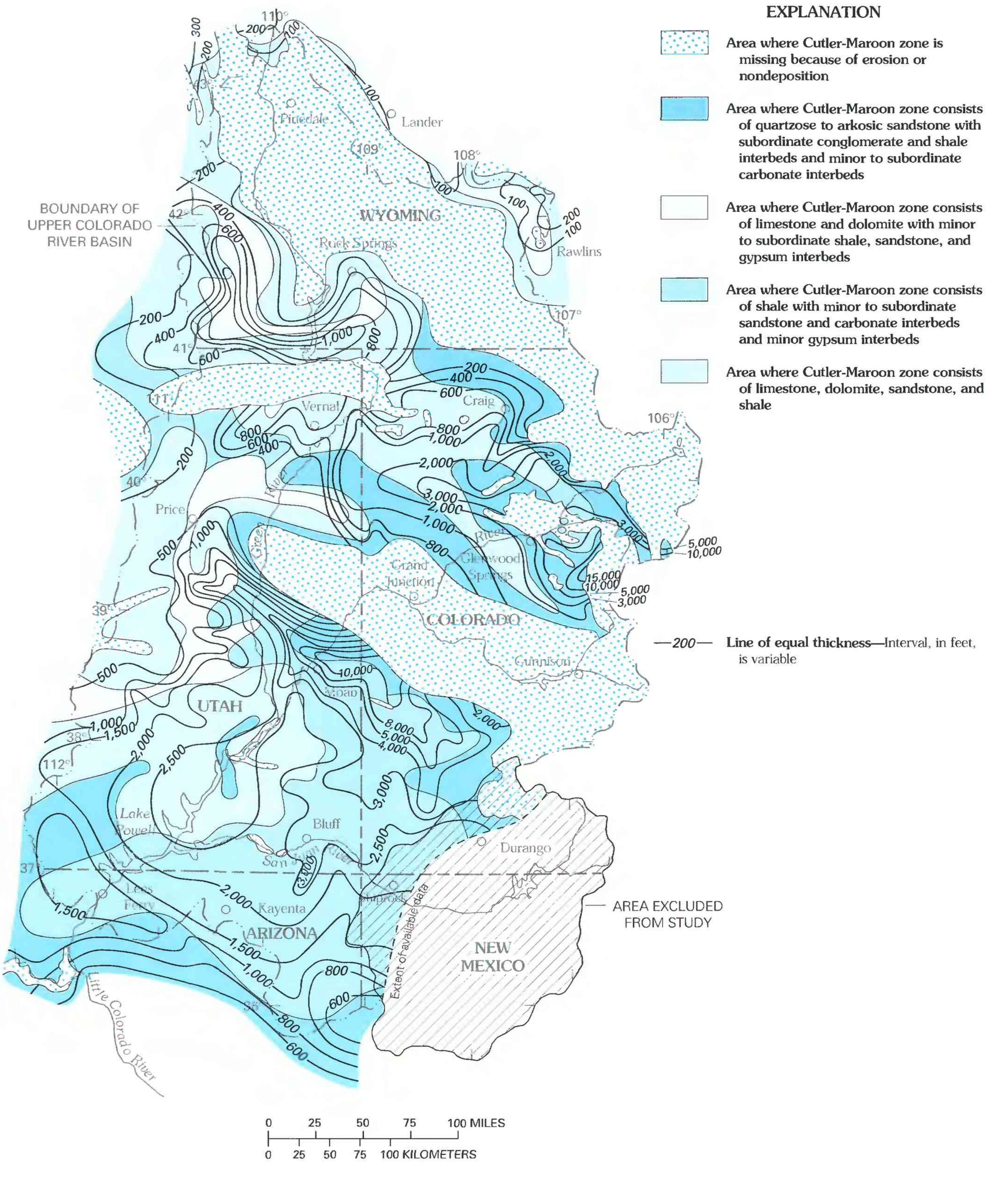


FIGURE 64.—Thickness and lithology of the Cutler-Maroon zone of the Canyonlands aquifer. (Modified from Geldon, in press, pl. 15.)

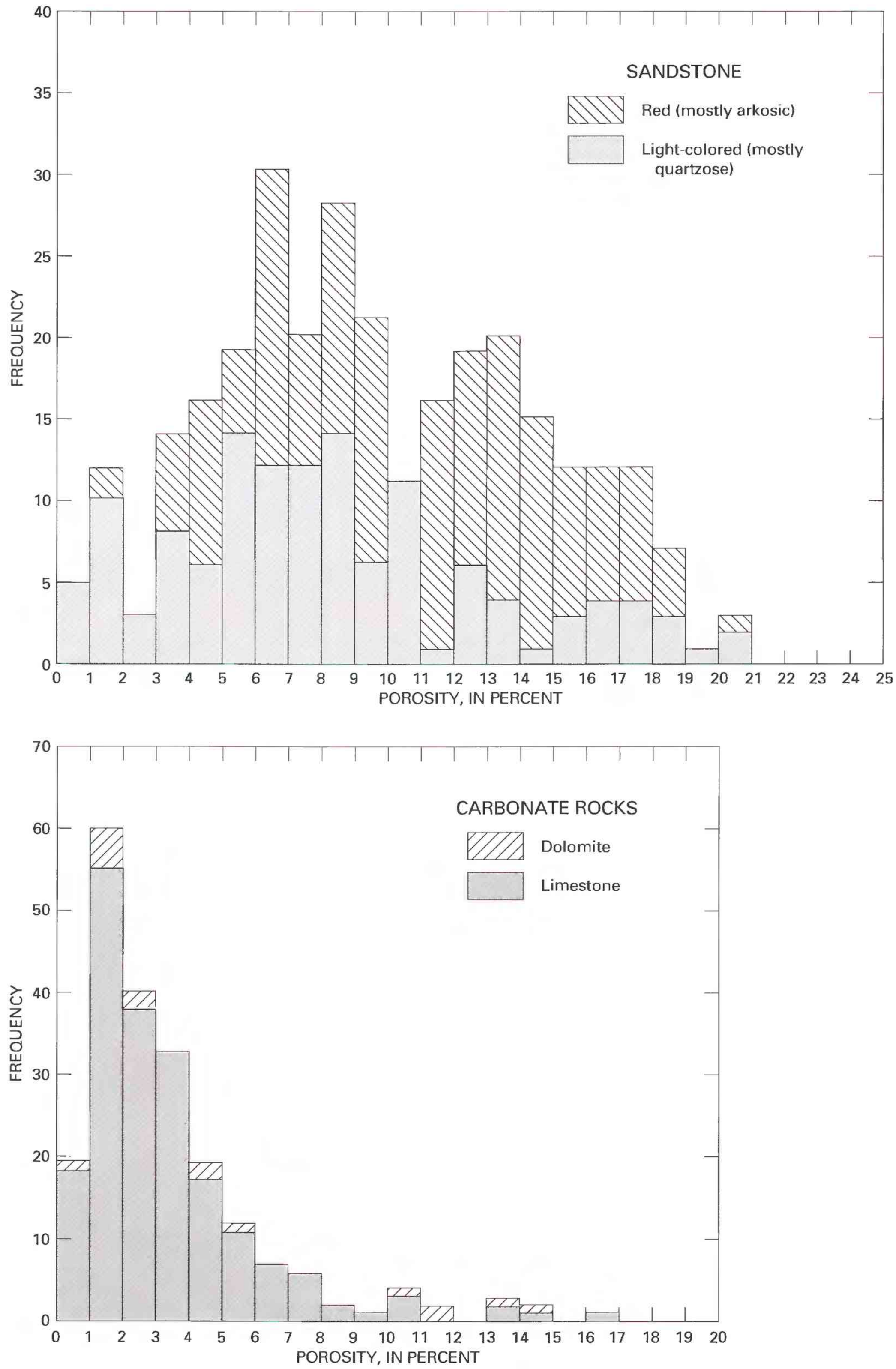


FIGURE 65.—Frequency distribution of porosity in samples of sandstone and carbonate rocks from the Cutler-Maroon zone of the Canyonlands aquifer.

TABLE 12.—*Porosity and pore-scale permeability statistics for the Cutler-Maroon zone of the Canyonlands aquifer*
[<, less than]

Rock type	Porosity (percent) Minimum	Porosity (percent) Maximum	Porosity (percent) Median	Number of observations	Pore-scale permeability (millidarcies) Minimum	Pore-scale permeability (millidarcies) Maximum	Pore-scale permeability (millidarcies) Median	Number of observations
Sandstone								
Limy	1.2	6.7	5.0	8	<0.01	<0.01	<0.01	8
Shaly	1.2	20.4	6.2	56	<.01	5.2	<.01	56
Friable	5.8	18.2	12.1	11	<.01	34	2.5	11
Silty	.8	17.9	4.5	51	<.01	8.9	.09	51
Fine grained	1.2	20.4	6.4	62	<.01	27	<.01	62
Medium grained	5.1	14.9	9.2	32	<.01	5.2	<.1	32
Coarse grained	6.3	18.6	15.4	27	.02	232	6.0	27
Gravelly	9.2	18.9	13.7	21	<.1	45	6.7	21
Mostly arkosic (red-colored)	1.4	20.4	11.6	153	<.01	232	2.0	153
Mostly quartzose (light-colored)	.8	20.0	7.3	145	.00071	27	.01	145
All	.8	20.4	9.1	299	.00071	232	.46	299
Limestone	.4	16.0	2.4	194	<.01	120	.01	194
Dolomite	.7	14.0	3.3	16	<.01	.98	.04	16
Shale	.5	11.9	4.7	10	<.01	.30	<.01	10

in a borehole drilled into the Cutler Formation, variations in porosity reflect the alternation of arkosic sandstone and shale intervals. However, individual sandstone or shale intervals cannot be distinguished. In figure 66*B*, a geophysical log in a borehole drilled into the Cedar Mesa Sandstone, intervals of siltstone or siltstone and quartz sandstone are distinctly less porous than intervals of quartz sandstone. In figure 66*C*, a geophysical log in a borehole drilled through several formations of different lithology, formation tops are clearly visible on the basis of changes in porosity. A distinct increase in porosity occurs at the contact between the Organ Rock Shale and the Cedar Mesa Sandstone. A distinct decrease occurs at the contact between the Cedar Mesa Sandstone and the lithologically diverse Elephant Canyon Formation. Within the lower formation, intervals of quartz sandstone or sandstone with limestone clearly are more porous than intervals of sandstone and siltstone or sandstone and dolomite. Carbonate intervals are the least porous. Variations in the porosity of sandstone in the Cedar Mesa Sandstone and Elephant Canyon Formation probably are attributable to variations in grain size and the amount of carbonate or silica cement.

Like porosity, pore-scale permeability in the Cutler-Maroon zone varies with lithology (table 12). In 299 analyses, the pore-scale permeability of sandstone ranged from 0.00071 to 232 md and generally increased with decreasing cement and clay content and increasing grain size (table 12). In 210 analyses, the pore-scale permeability of carbonate rocks ranged from less than 0.01 to 120 md and, on the average, was about an order of magnitude smaller than for sandstone. In 10 analyses, the pore-scale permeability of shale ranged from less than 0.01 to 0.30 md and was much smaller, on the average, than for all other rock types. Although median values of pore-scale permeability increase in about the same manner as median values of porosity for varieties of sandstone and carbonate rocks, a relation between the two properties appears to exist only for sandstone (table 3). As seen in figure 67, even this relationship is vaguely defined.

Local-scale permeability in the Cutler-Maroon zone depends not only on lithology but also on structural setting. In 71 determinations, local-scale permeability ranged from 0.00078 to 68 md, with a median value of 1.5 md (fig. 68). Intervals composed entirely of either sandstone or carbonate rocks were found to be more permeable than intervals containing shale interbeds. No significant differences in local-scale permeability were detected between sandstone and carbonate intervals. This indicates that secondary openings, such as fractures and solution channels, enhance the relatively small pore-scale permeability of carbonate rocks. Because secondary openings increase and cementation decreases toward uplifted areas, local-scale permeability of the Cutler-Maroon zone increases from structural basins to uplifted areas (fig. 69). In the Paradox Basin, local-scale permeability also appears to be affected by the northwest-trending anticlines and grabens that characterize the area.

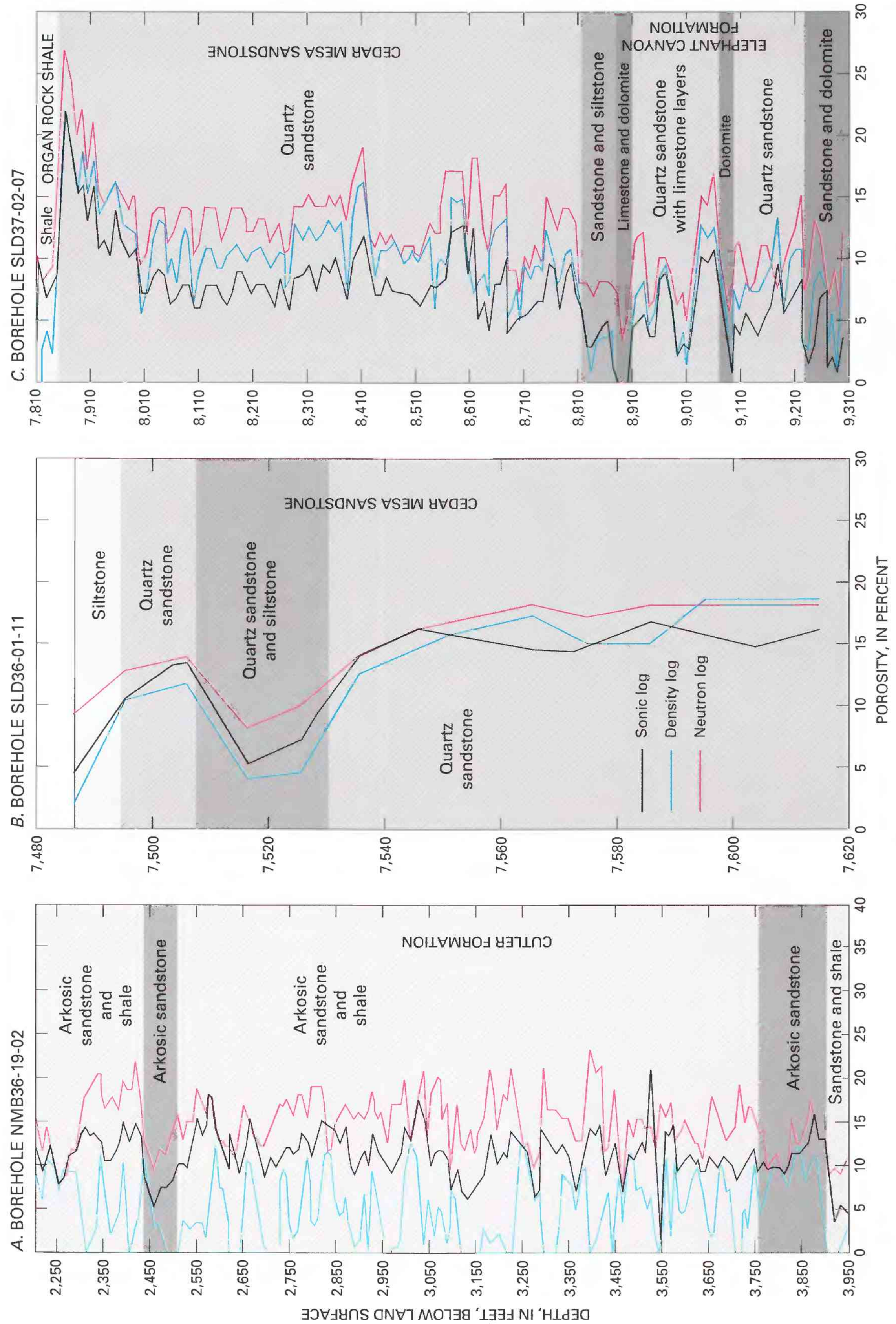


FIGURE 66.—Relation of geophysically determined porosity to depth below land surface and lithology in component geologic units of the Cutler-Maroon zone of the Canyonlands aquifer (stratigraphic nomenclature is that of Baars, 1962).

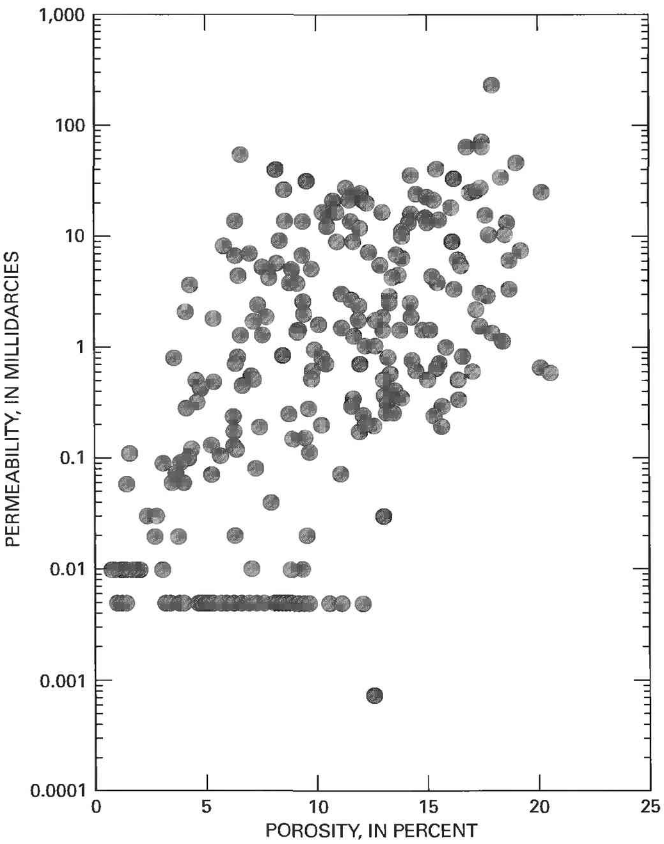


FIGURE 67.—Relation of porosity to pore-scale permeability in sandstone samples from the Cutler-Maroon zone of the Canyonlands aquifer.

HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

Hydraulic conductivity in the Cutler-Maroon zone, like local-scale permeability, ranges from small to large depending on lithology and structural setting. In 224 determinations (132 pressure-injection tests, 10 pumping tests, 6 bailing tests, 2 airlift tests, and 74 conversions from permeability or porosity determined by drill-stem tests, in the laboratory, or by geophysical methods), hydraulic-conductivity values ranged from 0.000002 to 10 ft/d (fig. 70), with a median value of 0.15 ft/d. Injection tests of the Maroon and Morgan Formations by the Bureau of Reclamation at four sites in northwestern Colorado indicated values of hydraulic conductivity that ranged from 0.037 to 4.6 ft/d (table 13). In these tests, sandstone intervals generally were more permeable than shale or limestone intervals, with many limestone and shale intervals taking in little or no water. Considerable variation in hydraulic conductivity among intervals with similar lithology may be related to the degree of fracturing, but there is no consistent variation in hydraulic conductivity with depth below land surface (fig. 71). The median hydraulic conductivity from 70 tests in five boreholes at the Ruedi Dam site, 0.76 ft/d, is very close to the hydraulic conductivity determined from a plot of residual drawdown with time during a bailing test of another borehole at the Ruedi Dam site. The data from this test (fig. 72) indicated the following:

$$\text{Discharge} = \frac{400\ \text{gal}}{87\ \text{min}} = 4.6\ \text{gal/min};$$

$$\text{Transmissivity} = \frac{35.2 \times 4.6\ \text{gal/min}}{(6.48 - 3.92)\text{ft}}$$

$$= 63\ \text{ft}^2/\text{d; and}$$

$$\text{Hydraulic conductivity} = (63\ \text{ft}^2/\text{d})/64\ \text{ft} = 0.98\ \text{ft/d}.$$

The injection and bailing tests, together, indicate an average hydraulic conductivity of about 0.9 ft/d for the 3,400-ft-thick section of the Cutler-Maroon zone at Ruedi Dam.

On the Monument Upwarp, four aquifer tests of the Cedar Mesa Sandstone and underlying formations composed of carbonate and clastic rocks indicated hydraulic-conductivity values ranging from 0.0011 to 0.027 ft/d (based on data reported by Huntoon, 1977, p. 19; Hand, 1979, p. 13; Hood and Danielson, 1981, p. 64–65; Woodward-Clyde Consultants, Inc., 1982, table 9–2; and Thackston and others, 1984, p. 26). However, an aquifer test of the Cedar Mesa Sandstone, alone, indicated a hydraulic-conductivity value of 2.5 ft/d (Richter, 1980, p. 18). The five aquifer tests on the Monument Upwarp, as in northwestern Colorado, indicate that shale and carbonate interbeds in the Cutler-Maroon zone have less potential to transmit water than sandstone.

Ranging from 0.001 to 10 ft/d, median values of hydraulic conductivity for sites on the Monument Upwarp and in northwestern Colorado define fairly well the range in unit-averaged hydraulic conductivity typical of uplifted areas. In contrast, drill-stem tests in structural basins generally indicate values of unit-averaged hydraulic conductivity ranging from 0.00005 to 0.001 ft/d (pl. 5). Only in the vicinity of fold axes, faults, and igneous intrusions do values of unit-averaged hydraulic conductivity in structural basins approach those recorded in uplifted areas. In the structurally complex Paradox Basin, for example, unit-averaged hydraulic-conductivity values of 0.003 to 0.02 ft/d have been obtained from drill-stem tests (pl. 5).

The composite transmissivity of the Cutler-Maroon zone conforms to the distribution of hydraulic conductivity, being larger in uplifted areas than in structural basins. Regionally, transmissivity is known from 13 widely scattered aquifer tests to range from 0.005 to about 900 ft^2/d (pl. 6). These 13 values and regional distributions of unit-averaged hydraulic conductivity and thickness indicate the composite transmissivity of the Cutler-Maroon zone probably ranges from less than 0.001 to greater than 1,000 ft^2/d (pl. 6). The Cutler-Maroon zone appears to be the

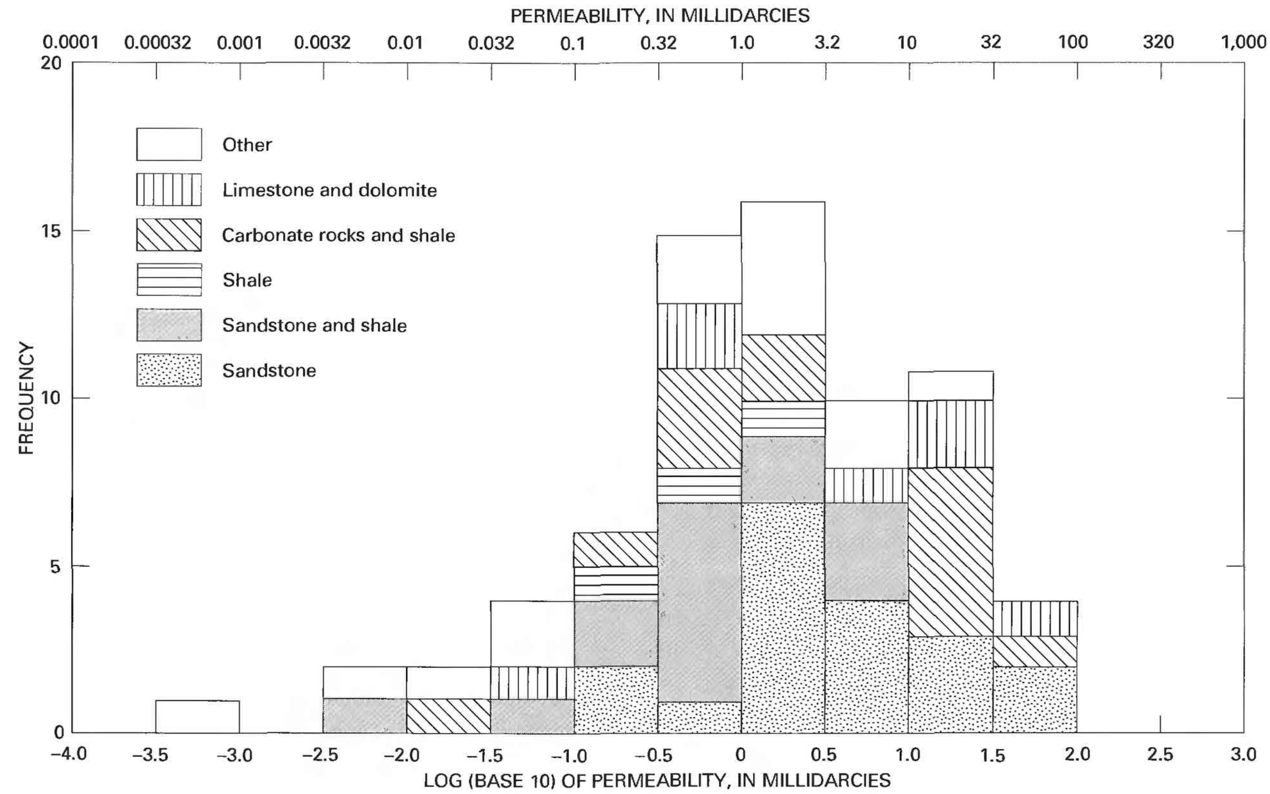


FIGURE 68.—Frequency distribution of local-scale permeability in the Cutler-Maroon zone of the Canyonlands aquifer.

most transmissive at the southern end of the Park Range, from the Sawatch Range and White River Plateau to the Axial Basin Arch and Uinta Mountains, in the Circle Cliffs Uplift the Abajo Mountains, and La Sal Mountains, and from the Monument Upwarp to the Chuska Mountains and Defiance Plateau.

YIELDS FROM WELLS AND SPRINGS

Yields from the Cutler-Maroon zone of the Canyonlands aquifer are small to moderate in most areas (pl. 6). In 175 determinations, artesian discharges from wells and springs ranged from 0.1 to 900 gal/min, with a median value of 6.7 gal/min (fig. 73).

The smallest discharges from the Cutler-Maroon zone consistently occur in structural basins; discharges from wells in these areas rarely exceed 10 gal/min. In the Paradox Basin, however, fractures associated with faulting, folding, and igneous intrusion locally have enhanced permeability, making yields from wells and springs of 10 to 300 gal/min possible. In the Henry Mountains and Blanding Basins and on the Four Corners Platform, fractures associated with igneous intrusions locally make yields of 10 to 50 gal/min possible from wells.

The largest discharges from the Cutler-Maroon zone consistently occur in uplifted areas, where fracturing, karst development, and weathering of cementing minerals from sandstone have enhanced permeability. In the Needles Fault Zone on the Monument Upwarp, for example, discharges from wells and springs, including Big Spring (SLD32–18–29dbd), commonly range from 50 to 125 gal/min (Huntoon, 1979, p. 43). On the White River Plateau and in the Elk and Uinta Mountains, springs with discharges of 50 to 100 gal/min, such as Dutch Spring (SC03–92–08bdc), Cold Iron Spring (SC10–88–04), and Burnt Springs (SB07–104–13bca), are not uncommon (Iorns and others, 1964; Sumsion, 1976; Teller and Welder, 1983), and springs with discharges of 400 to 900 gal/min, such as Stump Spring (SC03–92–23baa), Mitten Fault Spring (SB07–103–20dbc), and Arsenic Spring (SC11–87–35), can occur near faults (Rouse, 1967; Sumsion, 1976; Teller and Welder, 1983). The water from these large fault-controlled springs may be partly derived from Mississippian carbonate rocks affected by the faults. Numerous springs issuing from the Cutler-Maroon zone in Jones Hole (SLD03–25–01b) in the eastern Uinta Mountains, have a combined discharge of 16,650 gal/min (Hood, 1976, p. 34–35). However, these springs result more from draining of the entire Canyonlands aquifer into a topographic sink than from the transmissivity of the Cutler-Maroon zone alone. Under normal conditions, discharges from the Cutler-Maroon zone to wells and springs are not likely to exceed 900 gal/min.

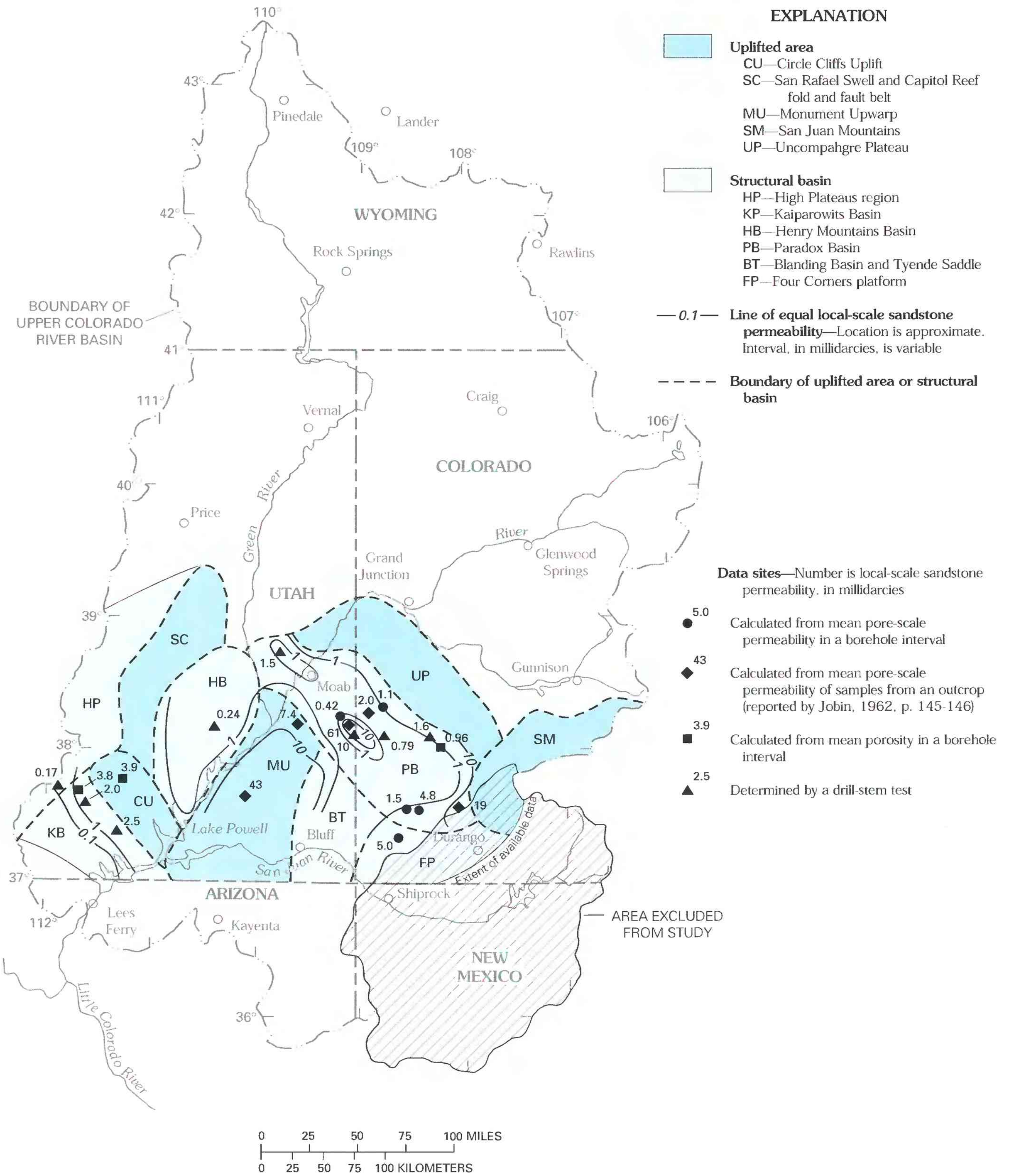


FIGURE 69.—Distribution of local-scale sandstone permeability in the Cutler-Maroon zone of the Canyonlands aquifer in southeastern Utah and southwestern Colorado.

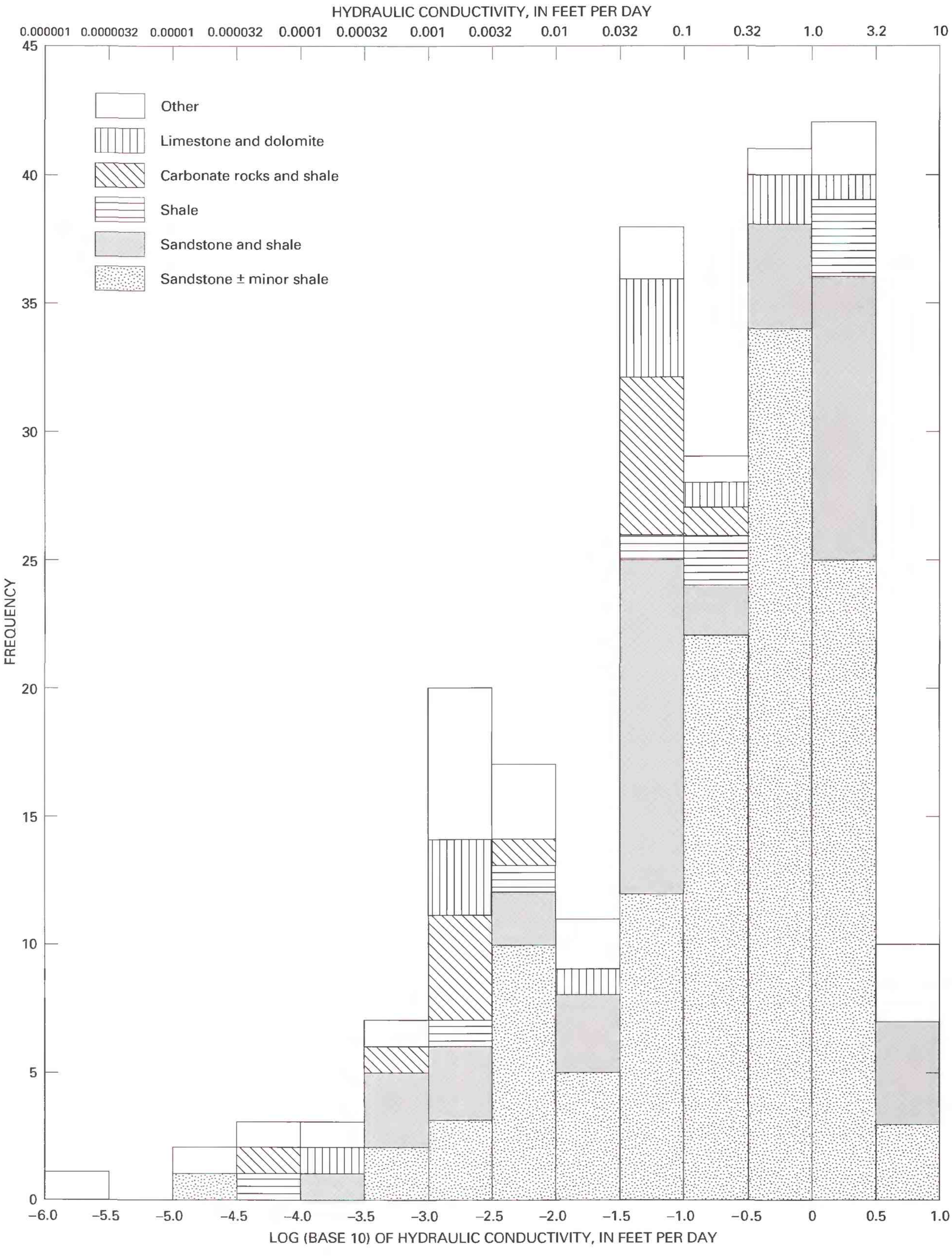


FIGURE 70.—Frequency distribution of hydraulic conductivity in the Cutler-Maroon zone of the Canyonlands aquifer.

TABLE 13.—*Values of hydraulic conductivity in the Cutler-Maroon zone of the Canyonlands aquifer determined by pressure-injection tests at dam sites in northwestern Colorado*

[Data are from Bureau of Reclamation, written commun., 1983–85]

Location	Borehole	Interval below land surface (feet)		Formation	Rock types	Hydraulic conductivity (feet per day)				Number of tests
		Top	Bottom			Minimum	Maximum	Median	Mean	
					Ruedi Dam site					
SC08–84–18bab	DH–30	2	300	Maroon	Sandstone with shale	0.12	4.6	0.86	1.1	19
SC08–84–18bad	DH–31	4	36	Maroon	Sandstone with shale	.60	3.0	.90	1.1	18
SC08–84–07cdc$_1$	DH–50	6	86	Maroon	Sandstone with shale	.21	4.2	1.0	1.5	5
SC08–84–07cdc$_2$	DH–51	2	100	Maroon	Sandstone with shale	1.3	4.4	3.0	2.9	4
SC08–84–18bab	DH–1000	35	1,001	Maroon	Sandstone with shale	.043	1.5	.062	.23	25
					Lake Avery Dam site					
SC01–91–18acb	DH–2	110	212	Maroon	Sandstone with siltstone	.072	.85	.20	.27	9
SC01–91–18bdd	DH–4	8	243	Maroon	Sandstone with siltstone	.081	.73	.25	.32	14
					Sawmill Mountain Dam site					
SB01–91–31bac$_1$	DH–1	90	260	Maroon	Sandstone with siltstone	.92	2.7	2.2	2.0	17
SB01–91–31bac$_3$	DH–2	110	150	Maroon	Sandstone with mudstone	1.8	3.7	2.9	2.8	8
					Juniper Mountain Dam site					
Sb06–94–18dba	DH–4	14	261	Morgan	Limestone and shale	.069	1.2	.20	.41	7
Sb06–94–18acd	DH–7	11	313	Morgan	Limestone, shale, and sandstone	.037	.94	.21	.41	6

WEBER-DE CHELLY ZONE

The Weber-De Chelly zone of the Canyonlands aquifer consists of the Wells Formation, Tensleep Sandstone, Weber Sandstone, White Rim Sandstone, De Chelly Sandstone, Coconino Sandstone, and the Schoolhouse and Fryingpan Members of the Maroon Formation (table 1). Component geologic units are Middle Pennsylvanian to Early Permian in age. Regionally composed mostly of well-sorted quartz sandstone, the Weber-De Chelly zone generally has at least some intergranular permeability and is capable of yielding small to moderate supplies of water to wells and springs nearly everywhere the zone is present. Component geologic units, such as the Tensleep Sandstone in the Rawlins Uplift (Berry, 1960), the Wells Formation in the Overthrust Belt (Lines and Glass, 1975), the Weber Sandstone in the eastern Uinta Mountains (Sumsion, 1976), and the De Chelly Sandstone in the Defiance Plateau and Chuska Mountains (Harshbarger and Repenning, 1954; Cooley and others, 1969), commonly are used to supply water to wells. However, these geologic units are so permeable that they may be nearly or entirely drained where they crop out in uplifted areas and are relatively thin (Metzger, 1961; Cooley and others, 1969; Sumsion and Bolke, 1972). In many of these uplifted areas, the only water supplies from the Weber-De Chelly zone are obtained from the bottom few feet, where downward-percolating meteoric water has ponded above less permeable shale or carbonate layers in the underlying Cutler-Maroon zone. The proportion of the UCRB where the Weber-De Chelly zone is unsaturated is relatively small and, therefore, the hydrogeologic unit can be expected to function as an aquifer throughout most of the UCRB.

THICKNESS AND LITHOLOGY

The thickness of the Weber-De Chelly zone ranges from 0 to about 4,000 ft (fig. 74). Over most of the UCRB, this hydrogeologic unit consists of light-colored quartz sandstone with generally thin interbeds of siltstone, mudstone, limestone, dolomite, and anhydrite that comprise no more than 10 percent of the thickness. However, carbonate rocks constitute 10 to 30 percent of the Weber-De Chelly zone on the western edge of the UCRB, from the Wasatch Plateau north to the Overthrust Belt, and in parts of the Green River and Great Divide Basins. Shale layers constitute as much as 30 percent of the Weber-De Chelly zone near its eastern and western depositional edges in Colorado, Utah, and New Mexico. Gypsum and anhydrite account for 5 to 10 percent of the Weber-De Chelly zone in a small area north and west of the Rock Springs Uplift. Contacts between component geologic units and overlying formations of Permian and Triassic age are gradational to unconformable.

POROSITY AND PERMEABILITY

Because it predominantly consists of well-sorted quartz sandstone, the Weber-De Chelly zone consistently is more porous than any other hydrogeologic unit in the UCRB. In 2,618 analyses, sandstone porosity ranged from less than 0.1 to

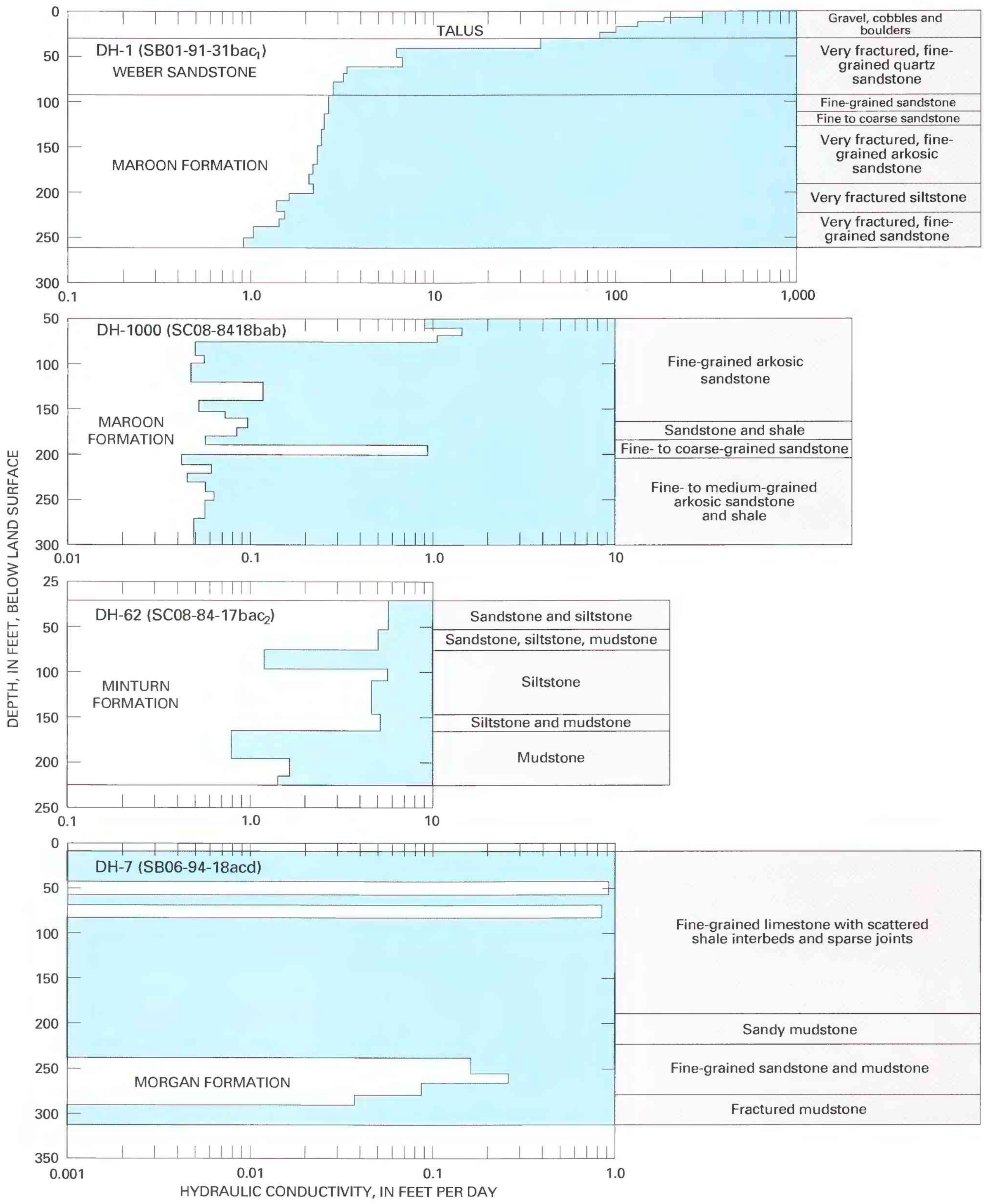


FIGURE 71.—Relation of hydraulic conductivity to depth below land surface and lithology in component geologic units of the Cutler-Maroon zone of the Canyonlands aquifer in northwestern Colorado (hydraulic-conductivity values from injection-test data furnished by Bureau of Reclamation, written commun., 1983–85).

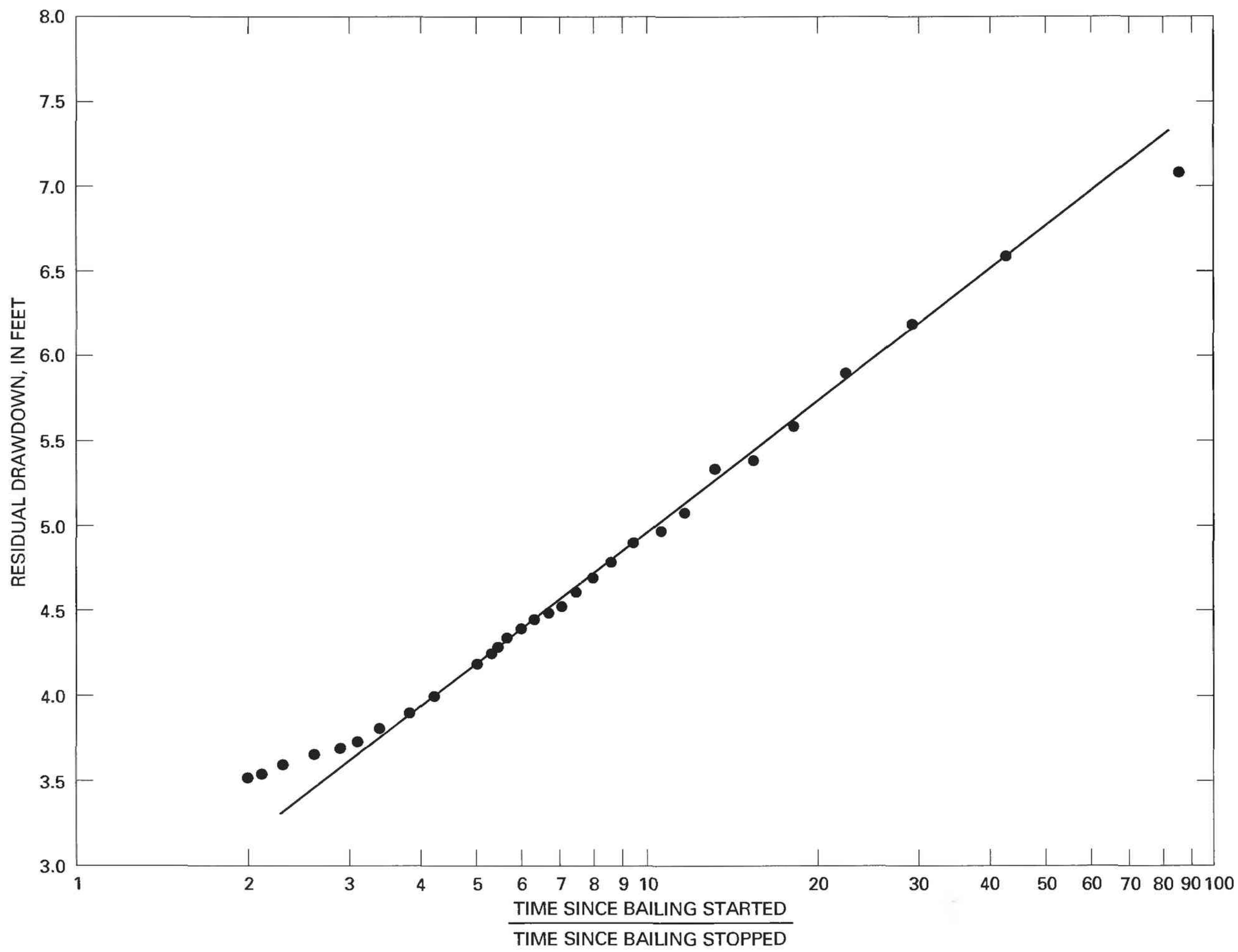


FIGURE 72.—Residual drawdown in well DH–16 (SC08–84–07bdc) at Ruedi Dam site, Colorado, during bailing test of Maroon Formation in April 1963 (data from Bureau of Reclamation, written commun., 1985).

27.7 percent, with a median value of 8.0 percent (fig. 75). Friable, micaceous, and shaly varieties were found to be more porous than indurated (limy or quartzitic) varieties; coarse-grained sandstone was found to be more porous than very fine-grained (silty), fine-grained, and medium-grained varieties (table 14). Though few in number, analyses of porosity in shale, carbonate rocks, and anhydrite indicate that these rock types, on the average, are less porous than sandstone in this hydrogeologic unit (table 14).

Borehole geophysical logs confirm that considerable variations in porosity can result from variations in sandstone texture and the presence of shale, carbonate, or evaporite interbeds. In figure 76*A*, for example, porosity in the Tensleep Sandstone at a site on the northern edge of the Great Divide Basin ranges from about 1 to 15 percent, probably reflecting variations in the amount of silica and carbonate cement; the relatively small porosity between depths of 4,810 and 4,850 ft probably indicates dolomite interbeds. In figure 76*B*, porosity in the Weber Sandstone at a site on the Axial Basin Arch ranges from about 3 to 15 percent, again probably reflecting variations in the degree of cementation and also the presence of siltstone and feldspathic sandstone interbeds. In figure 76*C*, porosity in the White Rim Sandstone at a site at the eastern edge of the San Rafael Swell ranges mostly from 11 to 15 percent, probably reflecting generally uniform grain size and cementation; intervals of small porosity between depths of 2,450 and 2,525 ft and at depths of about 2,900 and 3,025 ft probably indicate that carbonate or shale interbeds are present. In figure 76*D*, the porosity of the De Chelly Sandstone at a site on the western edge of the Monument Upwarp ranges from less than 1 to about 5 percent, indicative of considerable induration and, probably, interstitial clay and feldspar minerals that typically are present in this formation. Representative of vertical variations in porosity regionwide, these four examples emphasize the importance of sandstone texture in determining unit-averaged porosity.

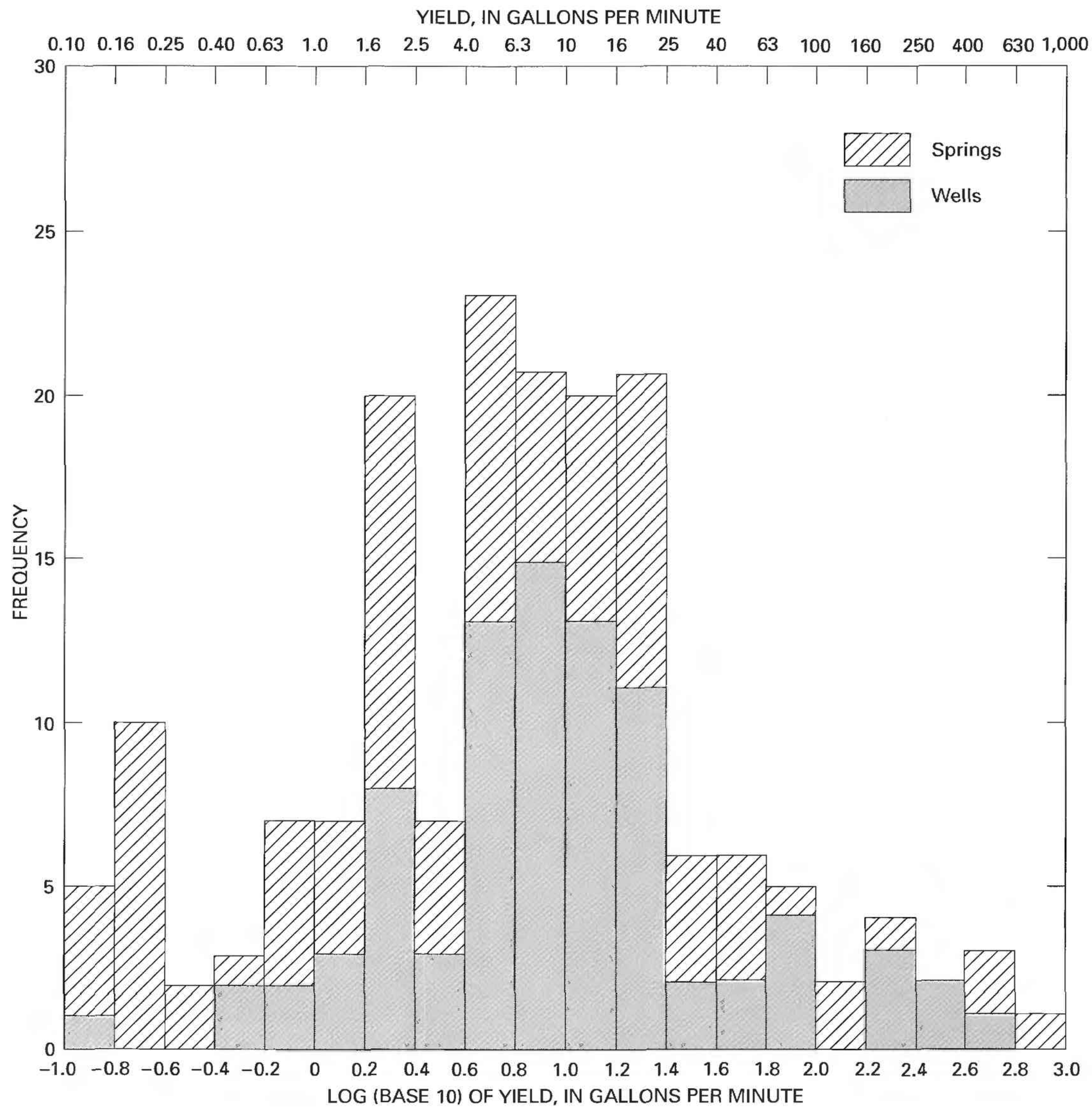


FIGURE 73.—Frequency distribution of yields from the Cutler-Maroon zone of the Canyonlands aquifer.

The largest values of unit-averaged porosity in the Weber-De Chelly zone occur in uplifted areas, where sandstone outcrops and subcrops commonly are leached of cementing minerals. In the Uinta Mountains, San Rafael Swell, Defiance Plateau, and San Francisco Plateau (a physical feature shown on pl. 1 that is beyond the political boundaries of the UCRB but within the boundaries of the ground-water system), the average porosity at outcrops of the Weber, White Rim, De Chelly, and Coconino Sandstones ranges from 11 to 28 percent (data from Cooley and others, 1969, p. 46–47, pl. 5; Hood, 1976, p. 31; and Hood and Patterson, 1984, p. 62–63). Unit-averaged porosity in this zone decreases substantially away from outcrop areas, as sandstone layers become more firmly cemented by silica and carbonate minerals precipitated from ground water (Bredehoeft, 1964; Fox and others, 1975). In the greater Green River Basin, the unit-averaged porosity of the Tensleep Sandstone, Weber Sandstone, and Wells Formation decreases from more than 15 to less than 5 percent toward the interior of the basin (Fox and others, 1975). On the basis of a large, regionwide distribution of geophysical and laboratory measurements of porosity and estimates of porosity from pore-scale permeability (using an equation given in table 3), it is estimated that unit-averaged porosity in the Weber-De Chelly zone ranges regionally from about 1 to 28 percent, increasing from structural basins to uplifted areas (pl. 7).

Permeability in the Weber-De Chelly zone also depends on lithology and structural setting and ranges from small to large. Pore-scale permeability in 2,618 analyses of sandstone ranged from less than 0.01 to 2.115 md, with a median value of 0.40 md (table 14). Local-scale permeability in 127 determinations

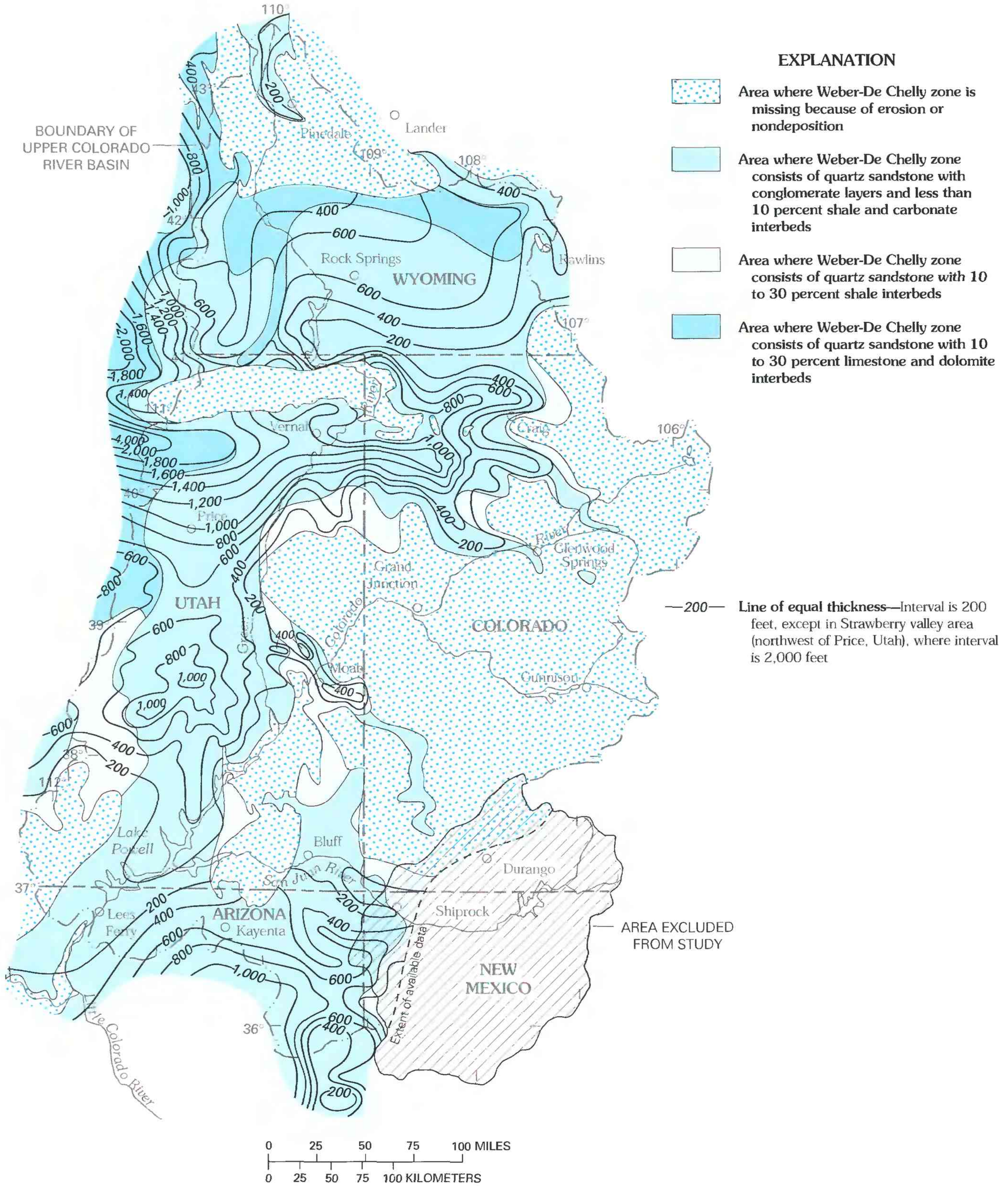


FIGURE 74.—Thickness and lithology of the Weber-De Chelly zone of the Canyonlands aquifer. (Modified from Geldon, in press, plate 16.)

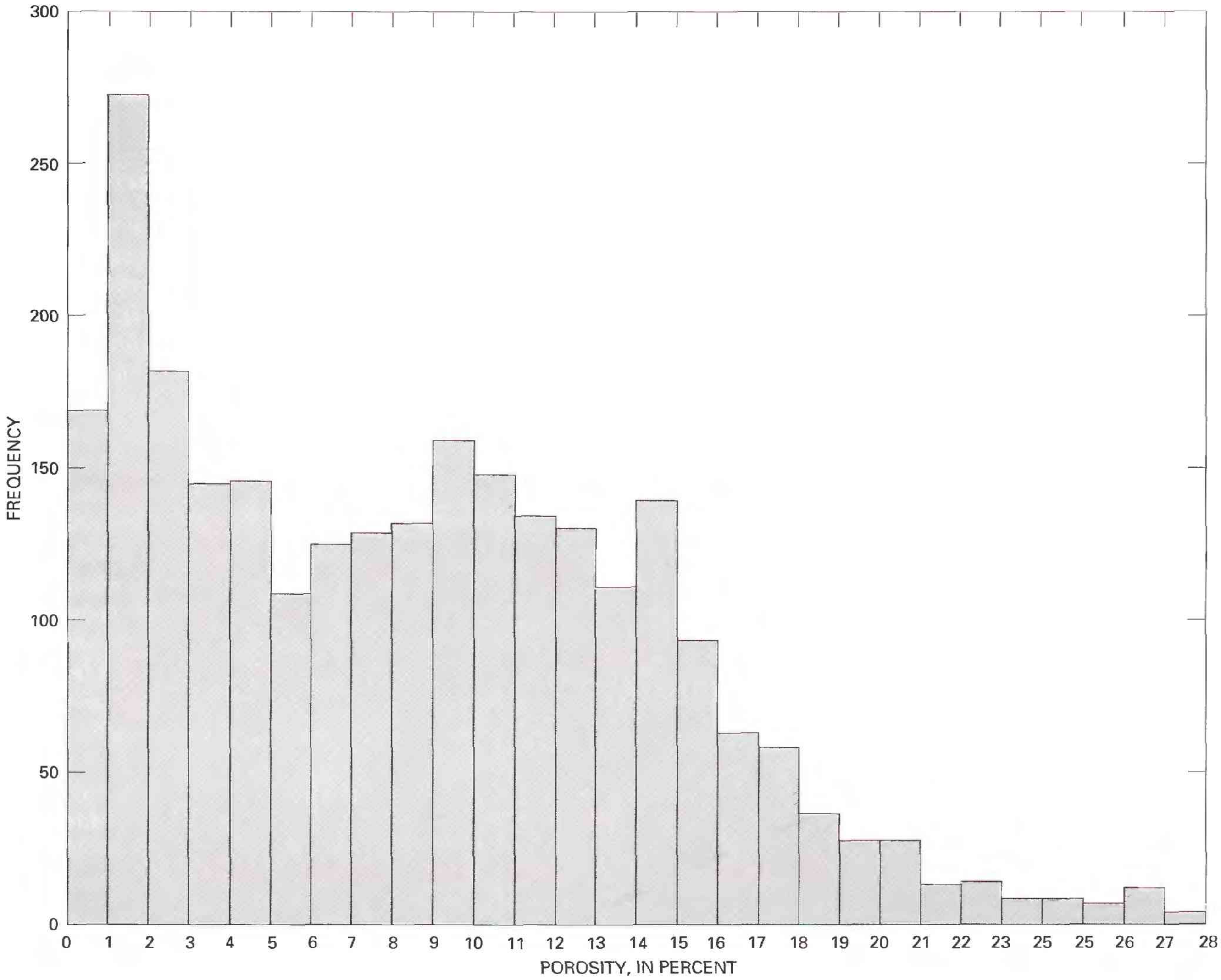


FIGURE 75.—Frequency distribution of porosity in samples of sandstone from the Weber-De Chelly zone of the Canyonlands aquifer.

ranged from 0.014 to 380 md, with a median value of 0.96 md (fig. 77). As indicated in table 14, friable and shaly varieties of sandstone, on the average, are more permeable than indurated (limy or quartzitic) varieties. Because grain size and cementation are two of the three factors that determine porosity (the other being sorting), it is not surprising that pore-scale permeability in the Weber-De Chelly zone is closely related to porosity (fig. 78 and table 3). Permeability variations not attributable to porosity probably can be explained by varying degrees of fracturing (Berry, 1960; Metzger, 1961; Akers and others, 1962; Whitcomb and Lowry, 1968; Sumsion and Bolke, 1972; Lines and Glass, 1975; Hood, 1976). Because fracturing increases and cementation decreases toward uplifted areas, the Weber-De Chelly zone generally is more permeable in uplifted areas than in structural basins (fig. 79).

HYDRAULIC CONDUCTIVITY, TRANSMISSIVITY, AND STORATIVITY

The hydraulic conductivity of the Weber-De Chelly zone commonly is small to moderate, but within and on the margins of uplifted areas (down hydraulic gradients from unsaturated outcrops), the hydraulic conductivity can be large. Hydraulic conductivity in 169 aquifer tests and calculations from local-scale permeability ranged from 0.000034 to 61 ft/d, with a median value of 0.012 ft/d (fig. 80). Values of hydraulic conductivity determined from pumping, flowing-well, and pressure-injection tests in and near uplifted areas ranged from 0.01 to 20 ft/d (table 15), whereas values determined from 40 drill-stem tests in basins or on the margins between basins and uplifts ranged from 0.000034 to 0.92 ft/d. On the basis of a large number of porosity, permeability, hydraulic conductivity, and transmissivity determinations, it is estimated that the unit-averaged hydraulic conductivity of the Weber-De Chelly zone ranges regionally from 0.00005 to 20 ft/d, increasing from structural basins to uplifted areas (pl. 7).

The composite transmissivity of the Weber-De Chelly zone, as indicated by aquifer tests listed in table 15 and regional distributions of thickness and unit-averaged hydraulic conductivity, is estimated to range from 0.01 to 6,000 ft^2/d (pl. 8). Values of less than 1 ft^2/d typically occur in structural basins, whereas aquifer tests in uplifted areas commonly indicate transmissivity values of 100 to more than 1,000 ft^2/d. The largest value of transmissivity in the Weber-De Chelly zone, 6,000 ft^2/d, was estimated from the specific capacity determined during a pumping test of the Weber Sandstone at the western end of the Uinta Mountains. That this value is at least the right order of magnitude is indicated by two conventionally analyzed aquifer tests of the Weber Sandstone in the eastern Uinta Mountains analyzed using standard methods. These tests indicated transmissivity values ranging from 1,400 to 4,000 ft^2/d.

The storativity of the Weber-De Chelly zone was estimated on the basis of regional distributions in thickness and unit-averaged porosity, using equation 22. Where the aquifer is more than 100 ft thick, the storativity was estimated to range from less than 0.0001 to 0.0013 (fig. 81).

YIELDS FROM WELLS AND SPRINGS

Yields from the Weber-De Chelly zone are small to moderate in most areas (pl. 8). In 177 determinations, artesian yields from wells and springs ranged from 0.1 to 2,200 gal/min, with a median value of 16 gal/min (fig. 82). With permeability reduced by cementation in structural basins, wells in these areas typically yield less than 25 gal/min. In uplifted areas where formations that comprise the Weber-De Chelly zone crop out, only the base of these formations may be saturated, and in these areas, springs issuing from the saturated zone typically have discharges of 0.1 to 5 gal/min. In uplifted areas where the component geologic units are buried, moderate to large discharges are possible. Olson Spring (SB23–88–20aaa), for example, issues from the Tensleep Sandstone in the Rawlins Uplift at a rate of 200 gal/min (Berry, 1960, p. 50–51). Echo Park well no. 3 (SB07–103–32adb) flows water from the Weber Sandstone in the eastern Uinta Mountains at rates of 35 to 150 gal/min (Sumsion, 1976, p. 42–45). Large springs, which may be related to nearby thrust faults, issue from the Wells Formation in the Overthrust Belt (Lines and Glass, 1975) and from the Weber Sandstone in the Uinta Mountains (Hood and others, 1976; Hood, 1977). Examples include Ratliff Spring (SLD02–22–31adc), with a discharge of 1,350 gal/min, Hams Fork Spring (SB29–118–13bad), with a discharge of 1,600 gal/min, and Warm Springs (UB01–08–30ddb), with a discharge of 2,200 gal/min.

Faults also appear to be responsible for discharges of 500 gal/min or more from the Weber-De Chelly zone on the flanks of the Wind River Mountains. Kendall Warm Spring (SB39–110–36) issues from the Phosphoria Formation on the northwestern flank of these mountains at a rate of about 2,900 gal/min. Because such large discharges are not typical

TABLE 14.—*Porosity and pore-scale permeability statistics for the Weber-De Chelly zone of the Canyonlands aquifer*

[<, less than]

Rock type	Porosity (percent)			Number of observations	Pore-scale permeability (millidarcies)			Number of observations
	Minimum	Maximum	Median		Minimum	Maximum	Median	
Sandstone								
Limy or quartzitic	<0.1	21.8	2.0	735	<0.01	970	0.03	735
Anhydritic	3.8	7.9	5.7	7	<.01	.30	<.01	7
Micaceous	4.9	16.2	8.3	27	<.01	3.2	<.01	27
Friable	2.7	14.6	9.4	36	<.01	30.0	1.8	36
Shaly		25.0	10.4	155	<.01	933	.88	155
Very fine grained	1.8	19.9	7.0	161	.01	35	.12	161
Fine grained	<.1	27.6	6.4	1,408	<.01	2,115	.15	1,408
Medium grained	.3	27.7	5.9	461	<.01	1,780	.10	461
Coarse grained to gravelly	.7	20.7	11.4	23	<.01	1,364	2.6	23
All	<.1	27.7	8.0	2,618	<.01	2,115	.40	2,618
Shale	4.6	11.2	6.8	8	<.01	.25	<.01	8
Limestone and dolomite	.6	6.0	1.4	8	<.01	.24	.01	8
Anhydrite	2.7	2.7	2.7	1	<.01	<.01	<.01	1

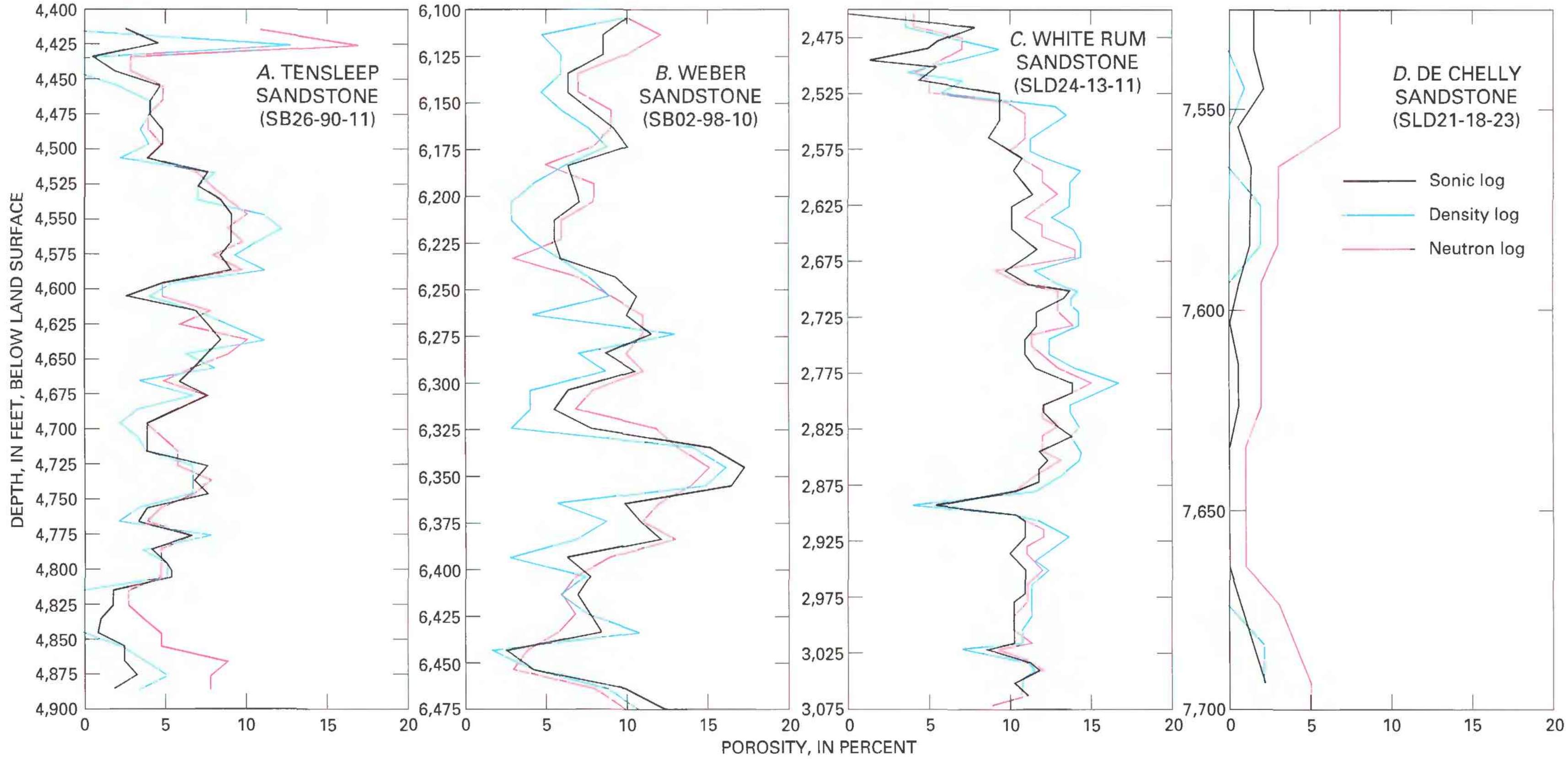


FIGURE 76.—Relation of geophysically determined porosity to depth below land surface in component geologic units of the Weber-De Chelly zone of the Canyonlands aquifer: *A*, Tensleep Sandstone; *B*, Weber Sandstone; *C*, White Rim Sandstone; *D*, De Chelly Sandstone.

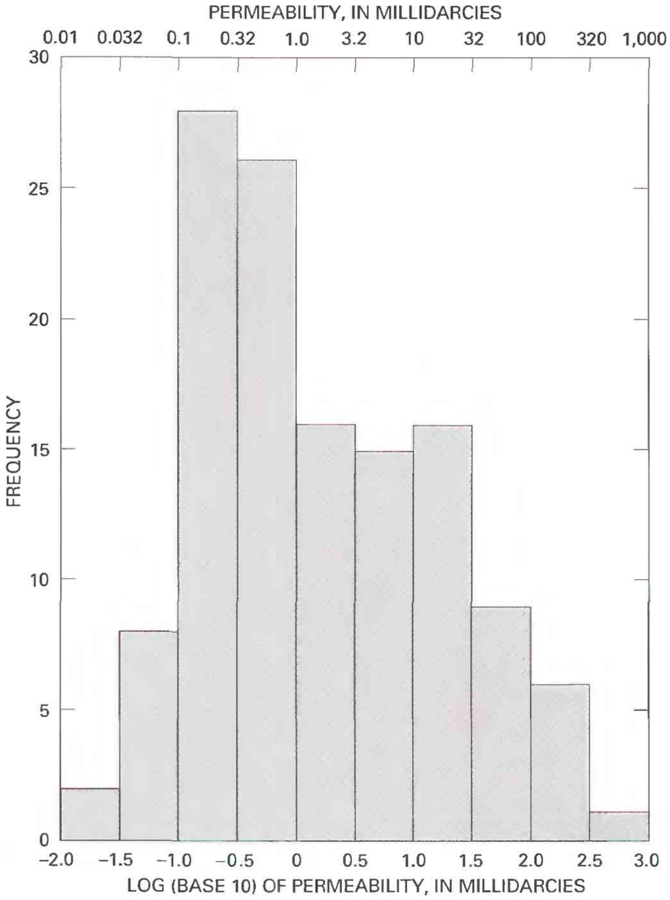


FIGURE 77.—Frequency distribution of local-scale permeability in the Weber-De Chelly zone of the Canyonlands aquifer.

of the Phosphoria Formation, it is presumed that most of the water issuing from this spring is produced by the disruption of ground-water circulation in the Tensleep Sandstone by the Wind River fault, a thrust fault that separates the Wind River Mountains from the Green River Basin. Kendall Warm Spring is situated right on this fault. Similarly, two flowing wells located on a thrust fault on the southeastern flank of the Wind River Mountains (SB33–99–35ad and SB33–100–25ca) also yield large quantities of water (500–540 gal/min) from the Tensleep Sandstone (Whitcomb and Lowry, 1968).

The largest discharges from the Weber-De Chelly zone in the UCRB occur from springs downstream from sinks in Mississippian carbonate rocks. Such springs, including Dry Fork Springs (SLD03–20–05c), Ashley Creek Springs (SLD03–20–01d), and Big Brush Creek Springs (SLD02–21–24c), discharge from the Weber Sandstone at seasonably variable rates that range from 500 to 9,000 gal/min (Maxwell and others, 1971; Hood and others, 1976). However, these discharges are more a consequence of topography than aquifer properties and are unlikely to be produced by wells or springs in typical hydrogeologic settings.

PARK CITY–STATE BRIDGE ZONE

The Park City–State Bridge zone of the Canyonlands aquifer consists entirely of Permian geologic units, including the Phosphoria, Park City, Kaibab, and Toroweap Formations and the lower parts of the Goose Egg and State Bridge Formations (table 1). Component geologic units generally have little intergranular permeability, but secondary permeability is imparted by fractures and solution channels. Consequently, small to moderate discharges to wells and springs from the Park City and Phosphoria Formations can occur in the Overthrust Belt (Lines and Glass, 1975) and Uinta Mountains (Hood and others, 1976), and small discharges are possible from the Toroweap and Kaibab Formations in the San Rafael Swell (Hood and Danielson, 1981), Capitol Reef Fold and Fault Belt (Marine, 1962), and northern Arizona (Metzger, 1961; Akers, 1964; Cooley and others, 1969). The State Bridge and Goose Egg Formations generally are not water bearing, but even these formations can yield small quantities of water to wells and springs from laterally persistent beds of sandstone, siltstone, or limestone (Berry, 1960; Brogden and Giles, 1976a; Giles and Brogden, 1976). Where the Park City–State Bridge zone is

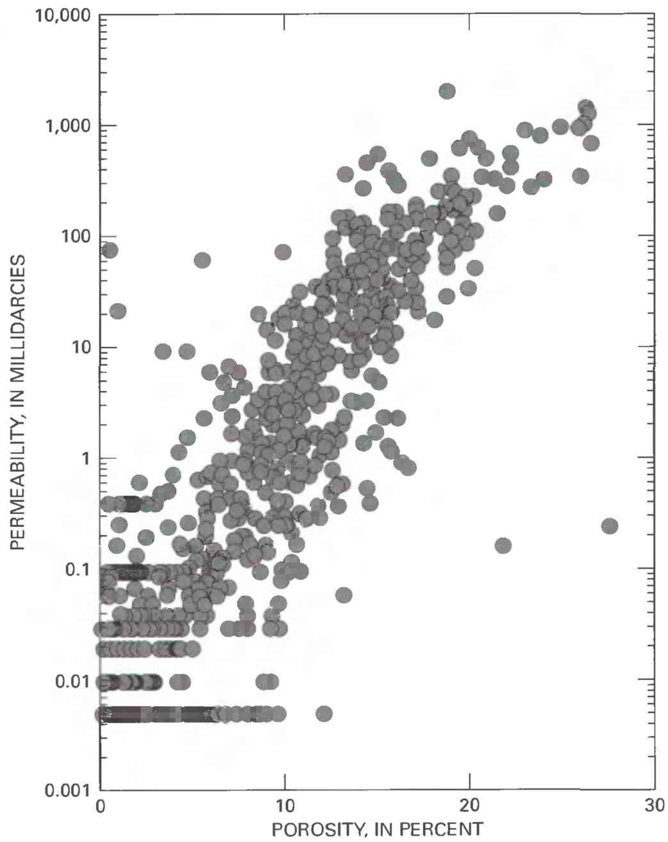


FIGURE 78.—Relation of porosity to pore-scale permeability in sandstone samples from the Weber-De Chelly zone of the Canyonlands aquifer.

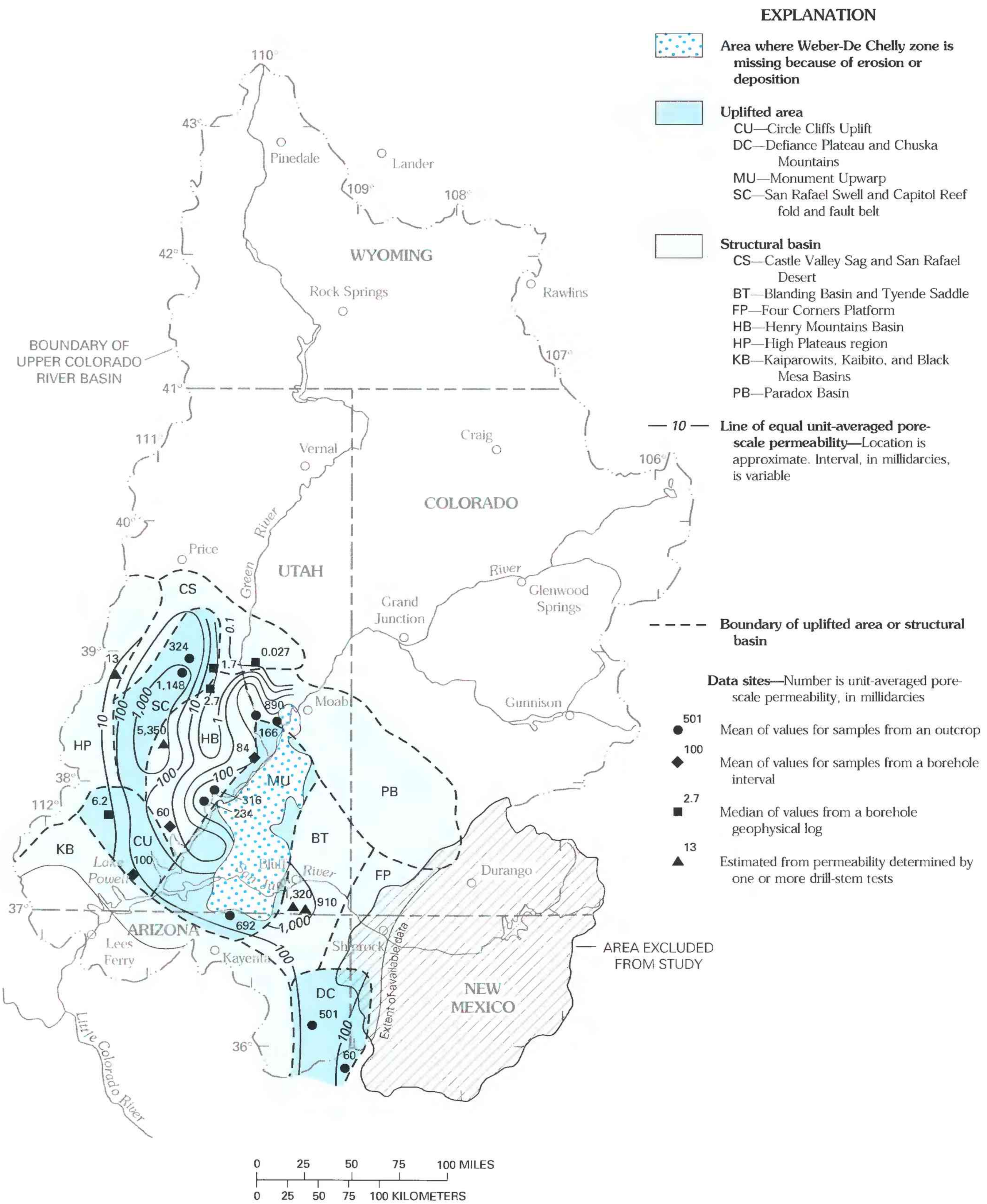


FIGURE 79.—Estimated distribution of unit-averaged pore-scale permeability in the Weber-De Chelly zone of the Canyonlands aquifer in southeastern Utah, northeastern Arizona, and northwestern New Mexico.

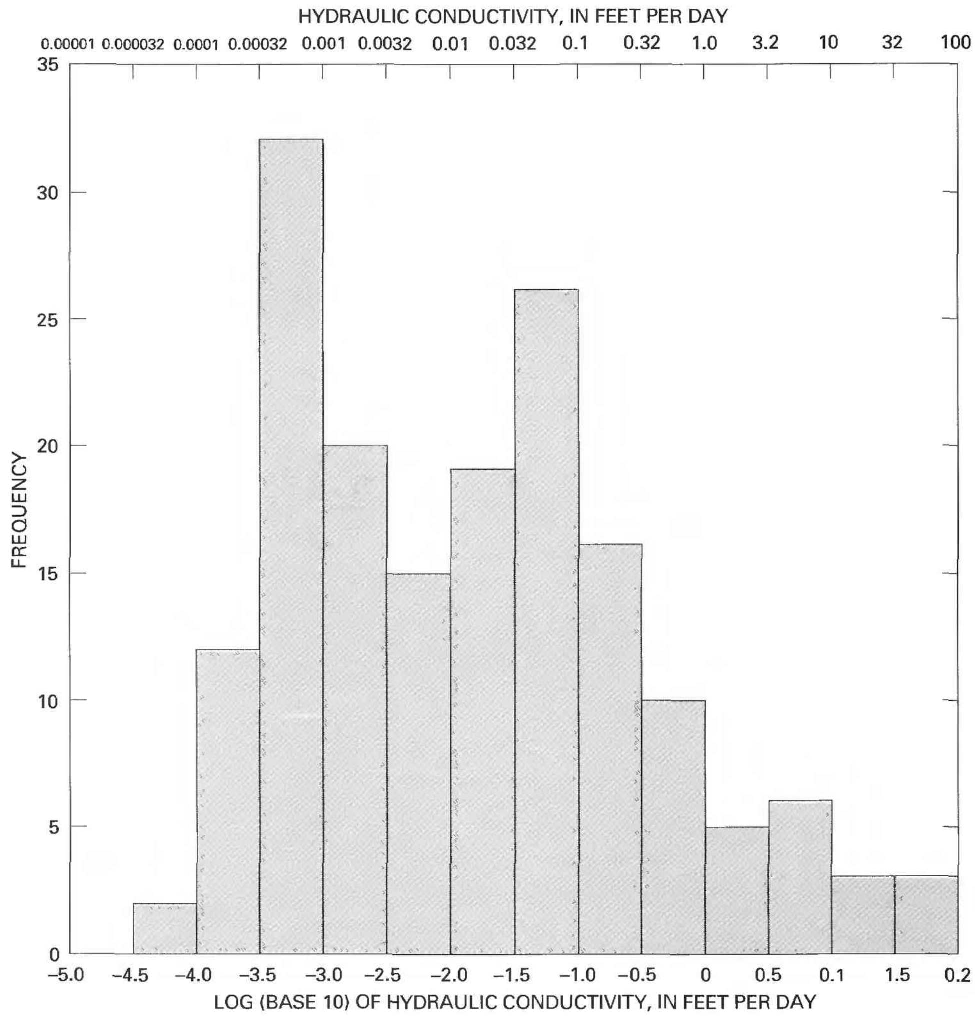


FIGURE 80.—Frequency distribution of hydraulic conductivity in the Weber-De Chelly zone of the Canyonlands aquifer.

water bearing, it generally is connected hydraulically to the underlying Weber-De Chelly zone. In northern Arizona, for example, wells commonly are completed in both the Kaibab Formation and Coconino Sandstone (Akers, 1964). Composed largely of shale and other rocks with little permeability over large parts of the UCRB and missing because of erosion or nondeposition in most of the area south and east of the confluence of the Green and White Rivers (fig. 83), the Park City–State Bridge zone can be expected to function as an aquifer subregionally in the UCRB.

THICKNESS AND LITHOLOGY

The thickness of the Park City–State Bridge zone ranges from 0 to 800 ft (fig. 83). In Wyoming and northwestern Colorado, a facies consisting of red shale with subordinate sandstone, gypsum-anhydrite, limestone, and dolomite grades westward into a depositional sequence consisting of varying proportions of limestone, dolomite, greenish-gray to black shale, phosphatic shale, phosphorite, chert, and sandstone. To the south, in Utah and Arizona, the shale and phosphatic intervals pinch out, and gypsum-anhydrite layers locally account for 5 to 30 percent of the Park City–State Bridge zone. Geologic units in the Park City–State Bridge zone are overlain conformably to unconformably by Triassic formations composed mostly of shale.

POROSITY AND PERMEABILITY

Large variations in porosity are exhibited by the diverse rock types that compose the Park City–State Bridge zone (fig. 84), but sandstone consistently is the most porous rock type (table 16). In 32 analyses, sandstone porosity ranged from 0.7 to 17.9 percent, with a median value of 7.4 percent. In 521 analyses, dolomite porosity ranged from less than 0.1 to 22.4 percent, with a median value of 2.1 percent. Median

TABLE 15.—*Values of hydraulic conductivity and transmissivity in the Weber-De Chelly zone of the Canyonlands aquifer determined from pumping, flowing-well, and injection tests in the Upper Colorado River Basin and vicinity*

[Dashes indicate not applicable; nd indicates no data; (A) indicates approximate]

Test site	Borehole		Type of test	Formation tested	Hydraulic conductivity (feet per day)			Transmissivity (feet squared per day)	Analytical method (if known)	Source of data
	Name	Location			Range	Site average	Area average			
Tensleep area[1]	Mills	SB44–87–08dcd	Flow	Tensleep Sandstone	nd	--		290	Estimate from specific capacity	Cooley (1985, p. 39)
Tensleep area	Davis	SB47–89–13aab	Flow	Tensleep Sandstone	nd	--	nd	150	Analysis of recovery data	do.
Tensleep area	Hamilton Ranch	SB50–90–14bbd	Flow	Tensleep Sandstone	nd	--		140		do.
Sawmill Mountain	DH–1	SB01–91–31bac1	Pressure-injection	Weber Sandstone	2.8–38	10		nd		Bureau of Reclamation (unpublished)
Sawmill Mountain	DH–2	SB01–91–31bac3	do.	Weber Sandstone	3.0–61	17	11	nd		do.
Sawmill Mountain	DH–3	SB01–91–31bac2	do.	Weber Sandstone	3.7–5.6	4.7		nd		do.
Echo Park	Well 3	SB07–103–32abd	Flow	Weber Sandstone	17	--	--	4,000	Analysis of recovery data	Sumsion (1976, p. 51–58)
Big Brush Creek	Stauffer Chemical	SLD02–22–29dcd	Pumping	Weber Sandstone	.1	--		1,400–2,700	do.	Hood (1976, p. 53–54)
Big Brush Creek	Stauffer Chemical	SLD02–22–32bcb	Pumping	Weber Sandstone	3.0	--	1.6	200	Estimate from specific capacity	do.
Kamas area[2]	Unnamed	SLD02–06–16cda	Pumping	Weber Sandstone	20	--	--	6,000	do.	Hood (1976, p. 54–56)
Taylor Canyon	Well 1	SLD27–17.5–01ddc	Flow	White Rim Sandstone	1.5	--		200	do.	Sumsion and Bolke (1972, p. 42)
Taylor Canyon	Well 2	SLD27–18–10aaa	Pumping	White Rim Sandstone	.19	--	.72	24	do.	do.
Taylor Canyon	Well 3	SLD27–18–09baa	Pumping	White Rim Sandstone	.48	--		80	do.	do.
Defiance Plateau	EPNG N–1, N–2	GA26–25–16	Pumping	De Chelly Sandstone	.27–.67	47		40–93	Estimate from specific capacity	Cooley and others (1969, p. 46–47)
Defiance Plateau	Ganado School	GA27–26–26c	Pumping	De Chelly Sandstone	.13	--		68	do.	do.
Defiance Plateau	Nazlini 2	NB02–09–16b	Pumping	De Chelly Sandstone	.47	--		89	do.	do.
Defiance Plateau	10T–239	NB03–10–15	Pumping	De Chelly Sandstone	.20	--	.35	100	do.	do.
Defiance Plateau	10T–272	NB05–10–20	Pumping	De Chelly Sandstone	.067	--		24	do.	do.
Defiance Plateau	17 wells	--	Pumping	De Chelly Sandstone	.013–3.5	--		7.5–490	do.	Davis and others (1963); McGavock and others (1966)
Leupp area[3]	Unnamed	GA22–12–12(A)	Pumping	Coconino	nd	--	--	4,700		Cooley and others (1969, p. 46–47)
St. John's area[4]	237–1	GA12–28–18(A)	Pumping	Coconino Sandstone and Kaibab Formation	nd	--	--	3,200	Computed from drawdown data	Akers (1964, p. 67)

[1]In Bighorn Basin, about 100 to 130 mi north of Upper Colorado River Basin (location of area shown on pl. 1).
[2]In Wasatch Range, about 16 mi west of Upper Colorado River Basin (location of area shown on pl. 1).
[3]In Lower Colorado River Basin, about 84 mi south of Upper Colorado River Basin (location of area shown on pl. 1).
[4]In Lower Colorado River Basin, about 97 mi south of Upper Colorado River Basin (location of area shown on pl. 1).

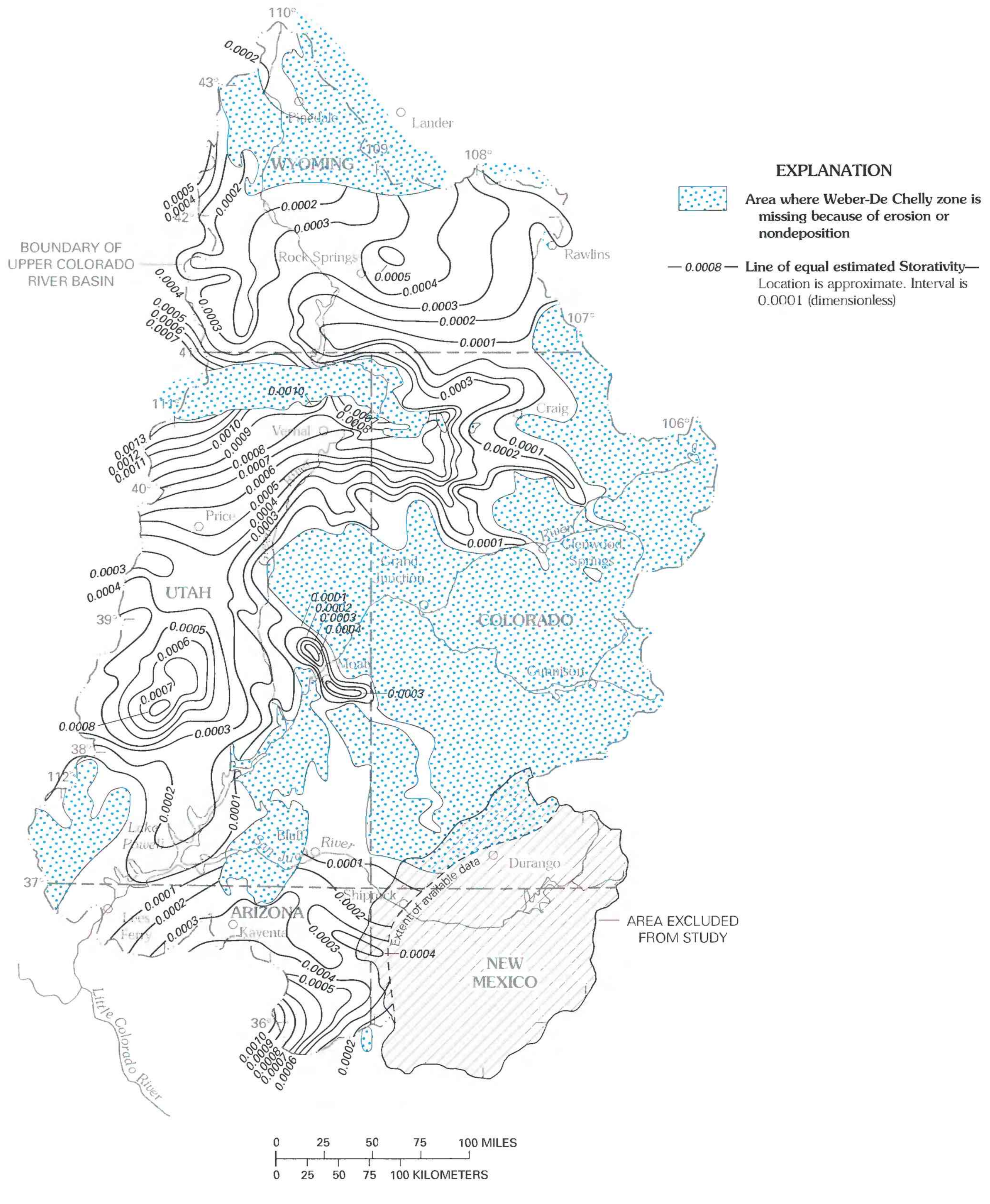


FIGURE 81.—Estimated storativity distribution in the Weber-De Chelly zone of the Canyonlands aquifer.

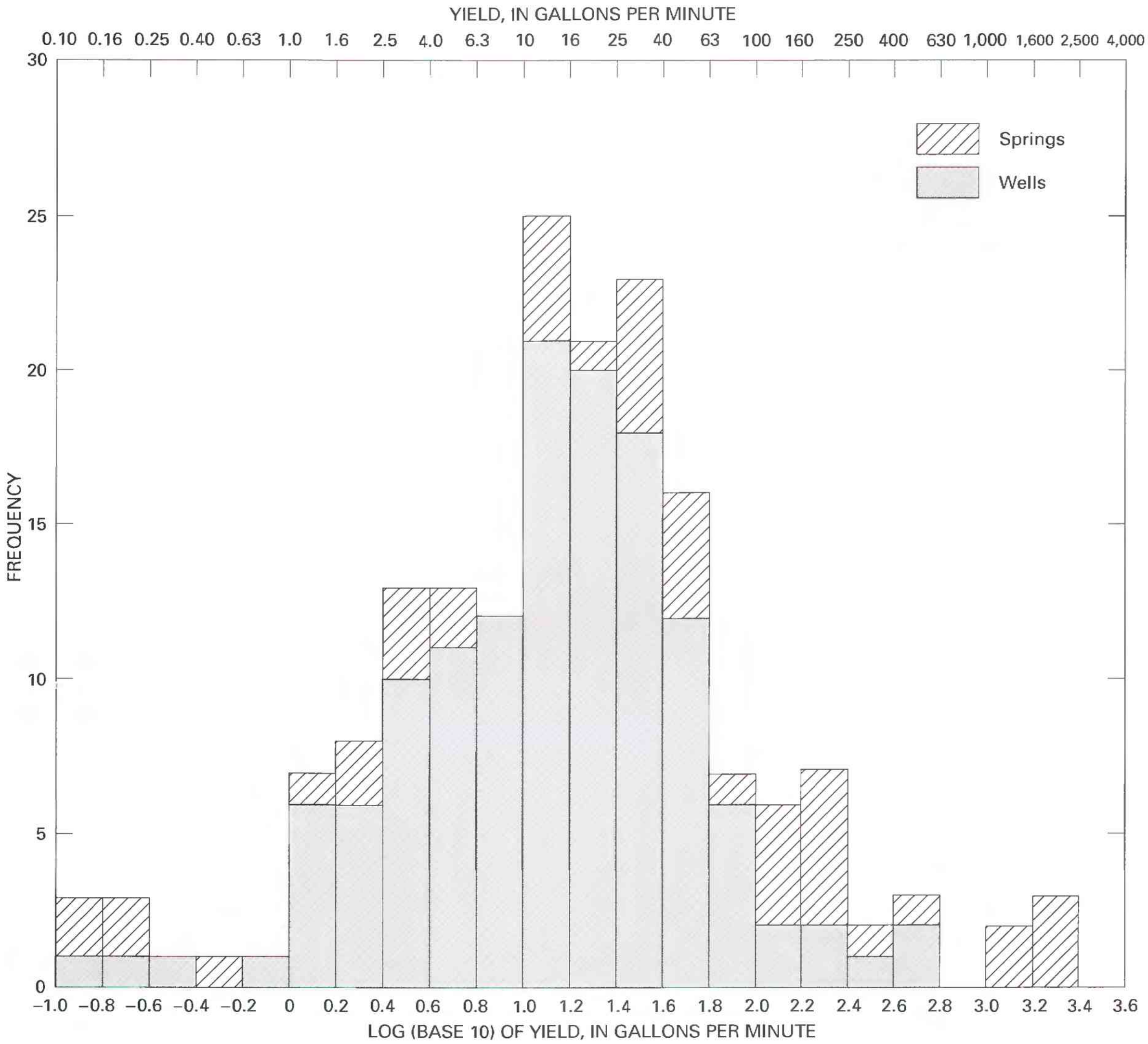


FIGURE 82.—Frequency distribution of yields from the Weber-De Chelly zone of the Canyonlands aquifer.

porosity values for most varieties of dolomite ranged from 1 to 2 percent, but the median porosity was about 4 percent in shaly dolomite and vuggy dolomite and about 10 percent in sandy dolomite. Limestone is somewhat more porous than dolomite on the average, although maximum values of limestone porosity are smaller than maximum values of dolomite porosity. In 76 analyses, limestone porosity ranged from 1.2 to 9.7 percent, with a median value of 3.4 percent. On the basis of three analyses, the porosity of anhydrite and shale appear to be consistent with values obtained for these rock types in other hydrogeologic units composed of Paleozoic rocks in the UCRB. Thus, median anhydrite porosity is estimated to be less than 1 percent, and median shale porosity is estimated to be between 2 and 4 percent. From the distribution of rock types in this hydrogeologic unit, median porosity values for these rock types, and the mean of laboratory-determined porosity values from widely scattered boreholes, it is estimated that the unit-averaged porosity of the Park City–State Bridge zone ranges regionally from about 2 to 6 percent (pl. 9).

Permeability in the Park City–State Bridge zone ranges from small to large. Pore-scale permeability in 629 analyses of sandstone, dolomite, and limestone ranged from less than 0.01 to 1,450 md (fig. 85), with median values of 0.04 md for sandstone, 0.06 md for dolomite, and 0.01 md for limestone. For sandstone and dolomite, a crude relation exists between pore-scale permeability and porosity (fig. 86). Most varieties of dolomite have median values of pore-scale permeability between 0.01 and 0.03 md, but sandy and vuggy varieties are considerably more permeable than other varieties (table 16). Local-scale permeability in the Park City–State Bridge zone is significantly

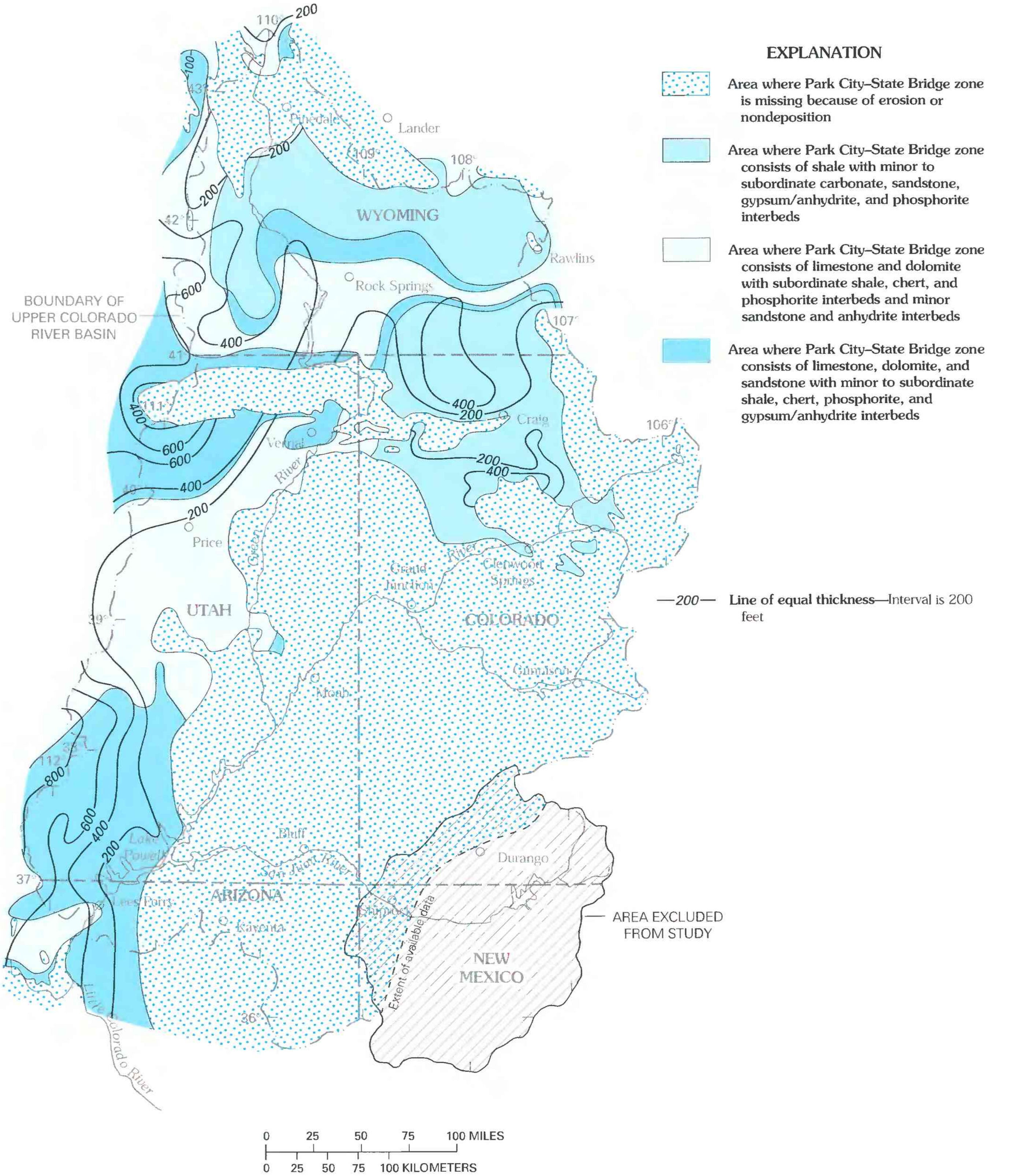


FIGURE 83.—Thickness and lithology of the Park City–State Bridge zone of the Canyonlands aquifer. (Modified from Geldon, in press, plate 17.)

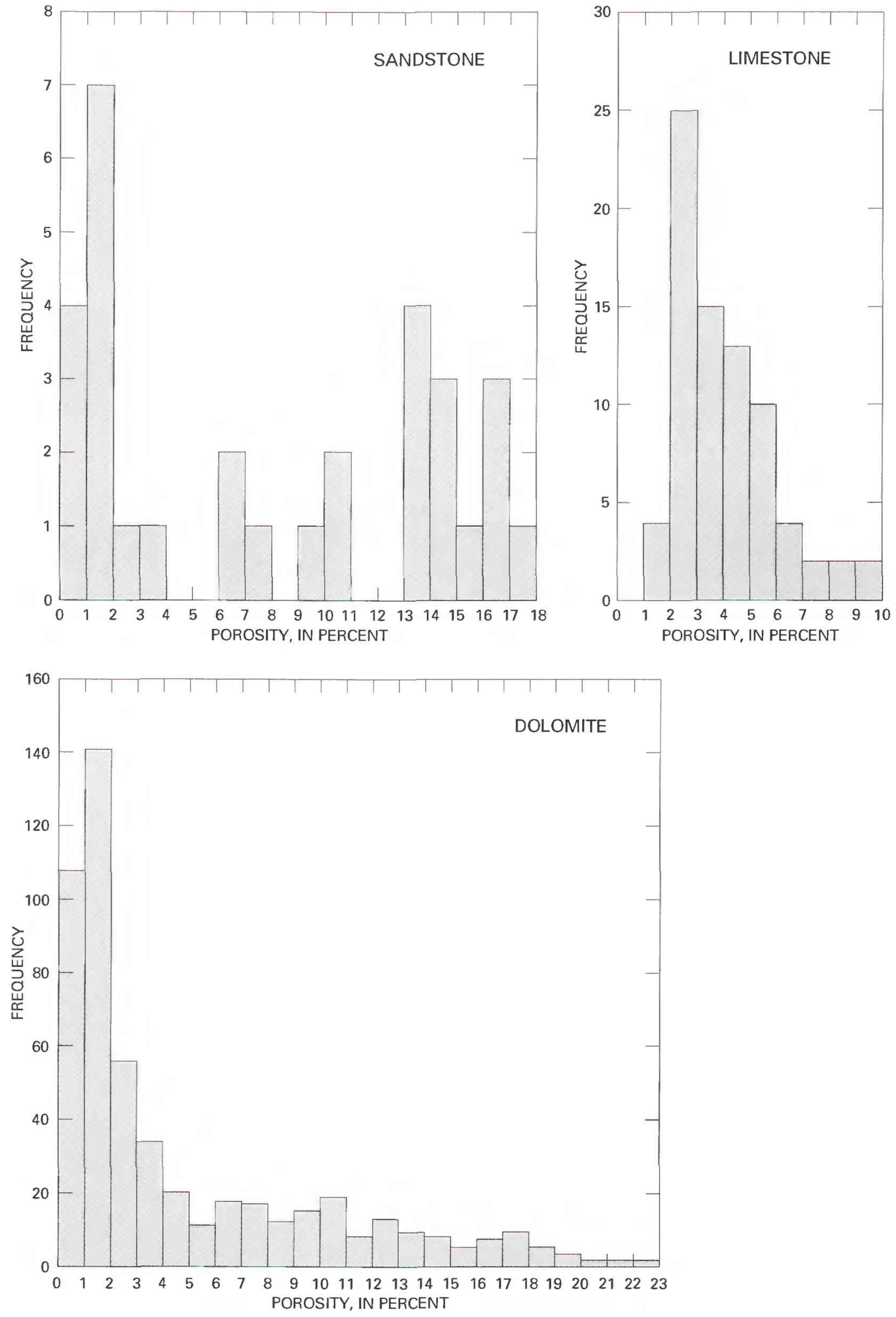


FIGURE 84.—Frequency distribution of porosity in samples of sandstone, dolomite, and limestone from the Park City–State Bridge zone of the Canyonlands aquifer.

TABLE 16.—*Porosity and pore-scale permeability statistics for the Park City–State Bridge zones of the Canyonlands aquifer*

[<, less than]

Rock type	Porosity (percent)			Number of observations	Pore-scale permeability (millidarcies)			Number of observations
	Minimum	Maximum	Median		Minimum	Maximum	Median	
Dolomite								
Phosphatic	0.2	10.5	1.1	42	0.01	46	0.03	42
Limy	.8	5.3	2.2	52	<.01	.50	<.01	52
Anhydritic	.7	14.7	2.2	50	<.01	21	.03	50
Vuggy	.7	12.8	3.9	33	.01	88	.07	33
Shaly	.5	6.9	4.3	15	<.01	21	.01	15
Fine-grained	.3	3.5	.9	39	<.01	23.6	.03	39
Crystalline	.2	21.8	1.5	303	<.01	88	.03	303
Sandy	.9	15.7	10.2	29	.03	36	.89	29
All	<.1	22.4	2.1	521	<.01	1,450	.06	521
Limestone	1.2	9.7	3.4	76	<.01	102	.01	76
Sandstone	.7	17.9	7.4	32	<.01	187	<.01	32
Shale	4.5	4.5	4.5	1	<.01	<.01	<.01	1
Anhydrite	.4	.4	.4	2	.01	.39	.20	2

larger than pore-scale permeability because of fracturing and solution. In 20 drill-stem tests and 15 conversions from pore-scale permeability, local-scale permeability in intervals composed mostly of limestone, dolomite, or sandstone ranged from 0.0049 to 330 md, with a median value of 2.4 md. Insufficient data were available to determine the range in permeability and median permeability values for shale, chert, phosphorite, or anhydrite, which comprise much of the Park City–State Bridge zone in northwestern Colorado and much of Wyoming.

HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

On the basis of the previously discussed permeability data, hydraulic conductivity in intervals of the Park City–State Bridge zone composed mostly of limestone, dolomite, or sandstone was determined to range from 0.000012 to 0.80 ft/d, with a median value of 0.006 ft/d (fig. 87). It is estimated that unit-averaged hydraulic conductivity in the center of the Green River, Great Divide, Washakie, and Sand Wash Basins, where the Park City–State Bridge zone is deeply buried and predominantly composed of shale, is no larger than the smallest recorded interval values. Toward the center of uplifted areas, where the Park City–State Bridge zone is fractured and predominantly composed of carbonate rocks and sandstone, unit-averaged hydraulic conductivity may be as large as 1 ft/d. Regionally, unit-averaged hydraulic conductivity is estimated to range from 0.00001 to 1 ft/d, increasing from structural basins to uplifted areas (pl. 9).

On the basis of regional distributions of unit-averaged hydraulic conductivity and thickness and an empirical relation between transmissivity and the discharge of wells and springs (discussed below), the composite transmissivity of the Park City–State Bridge zone is estimated to range from 0.005 to 70 ft^2/d (pl. 10). No estimates of composite transmissivity can be made for the least permeable geologic unit within this zone, the State Bridge Formation, but the transmissivity of the State Bridge Formation should be about the same as the lithologically similar Goose Egg Formation, which comprises the Park City–State Bridge zone along the eastern edge of the UCRB in Wyoming. In general, the composite transmissivity of the Park City–State Bridge zone increases from structural basins to uplifted areas.

YIELDS FROM WELLS AND SPRINGS

Yields from the Park City–State Bridge zone to wells and springs typically are small to moderate (pl. 10), but large yields are possible where composite transmissivity values exceed 50 ft^2/d. In 52 determinations, artesian yields from wells and springs ranged from 0.39 to 2,870 gal/min, with a median value of 13 gal/min (fig. 88). In general, yields, in gallons per minute, were found to be related to composite transmissivity, in feet squared per day, by a factor of 10 to 1. Thus, where composite transmissivity ranges from 0.01 to 1 ft^2/d, yields can be expected to range from 0.1 to 10 gal/min. Such yields typically are reported for drill-stem tests in structural basins or on the margins between basins and uplifts. Yields of 10 to 100 gal/min, which are typical of the Park City Formation in the Ashley Valley oil field (pl. 10), characterize discharges on the flanks of uplifted areas.

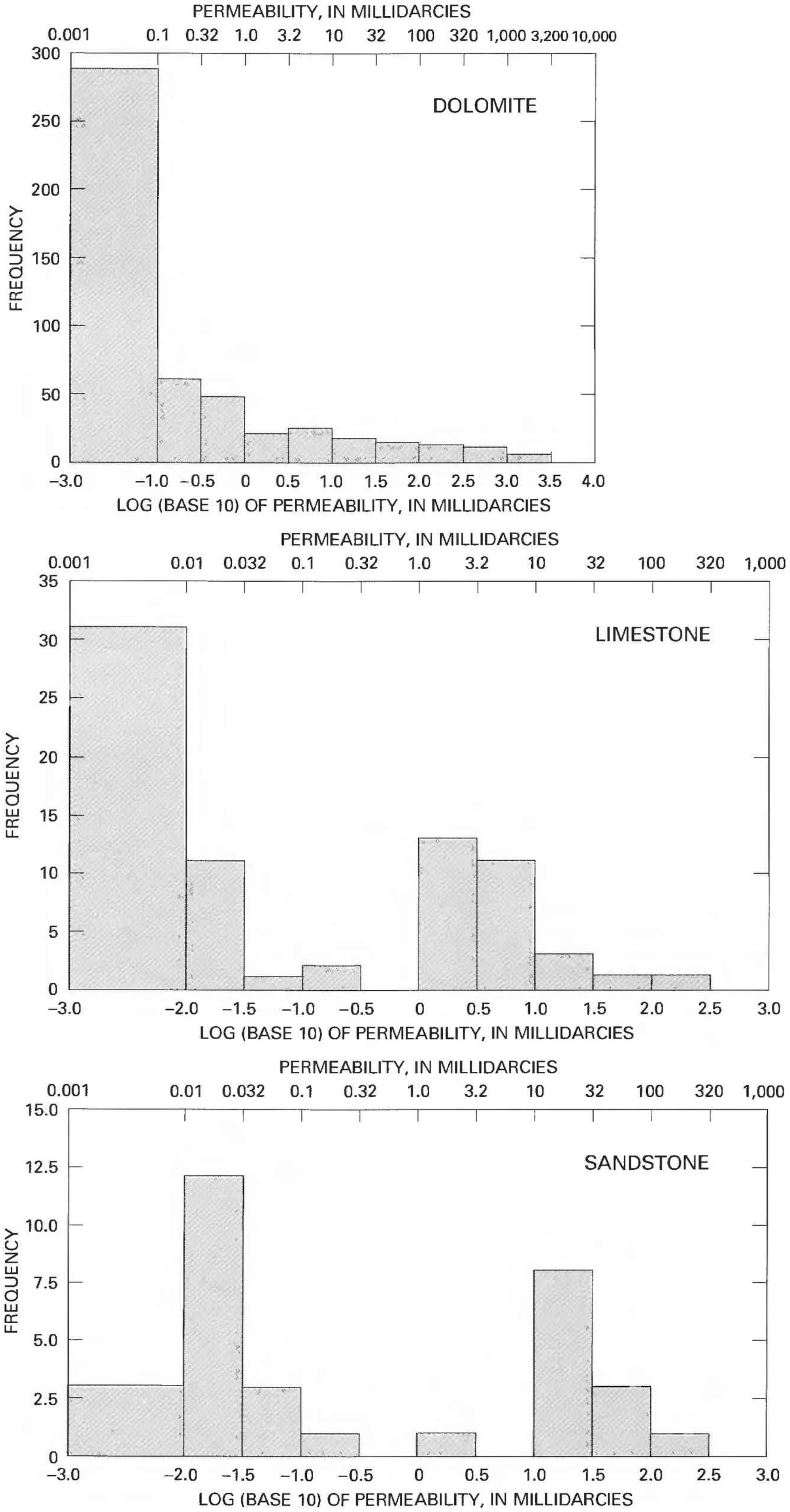


FIGURE 85.—Frequency distribution of pore-scale permeability in the Park City–State Bridge zone of the Canyonlands aquifer.

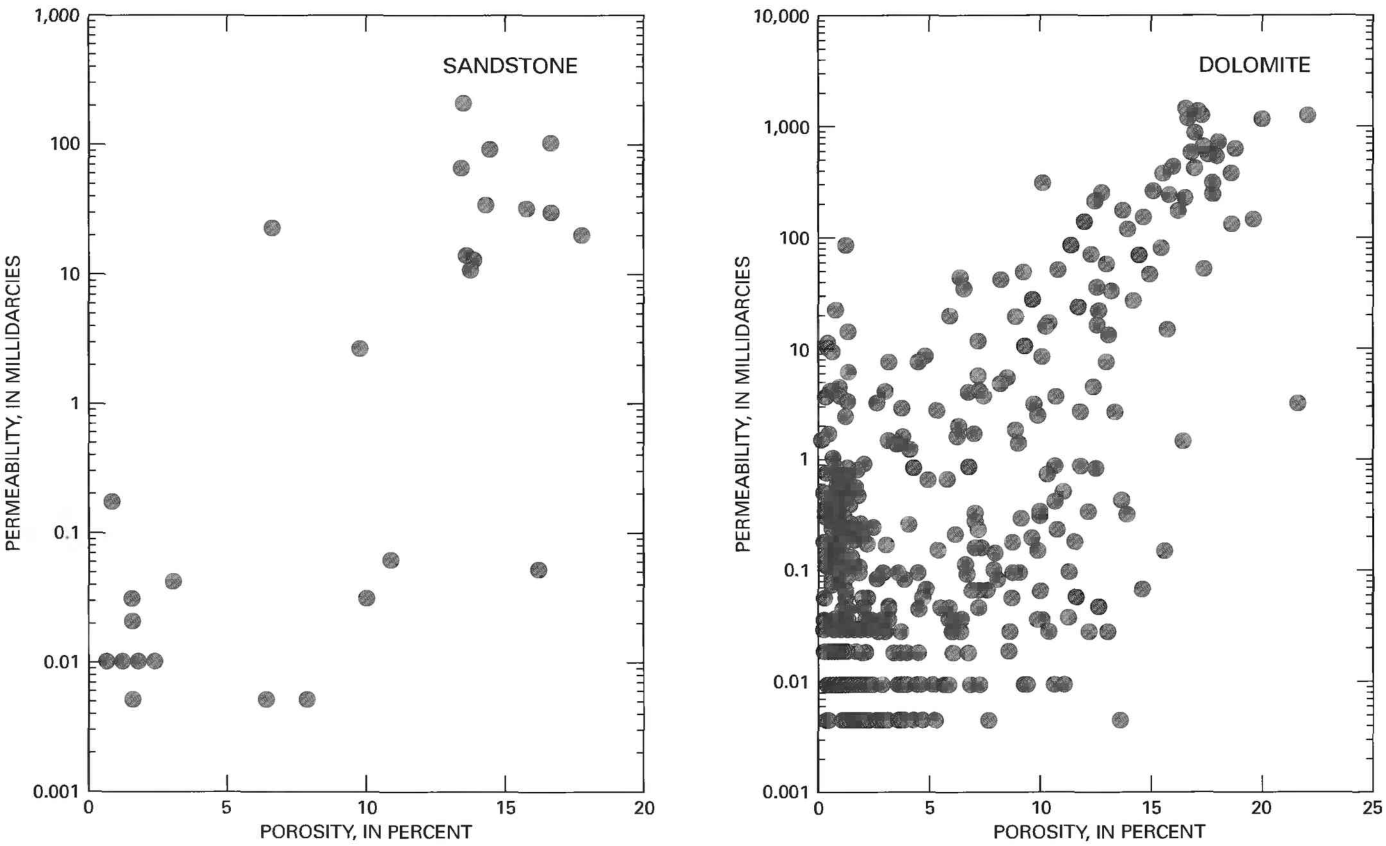


FIGURE 86.—Relation of porosity to pore-scale permeability in samples of sandstone and dolomite from the Park City–State Bridge zone of the Canyonlands aquifer.

The largest discharges from the Park City–State Bridge zone occur in uplifted areas, where the rocks are most fractured. As an example, springs and flowing wells near Kemmerer, Wyo., in the Overthrust Belt, discharge from the Phosphoria Formation at rates of 200 to 300 gal/min (Lines and Glass, 1975). At the eastern end of the Uinta Mountains, Taylor Ranch Warm Spring (SLD02–22–24ccd) discharges from the Park City Formation at rates varying between 112 and 449 gal/min (Maxwell and others, 1971, p. 21). Such discharges probably represent the maximum possible at the eastern end of the Uinta Mountains, where the Park City–State Bridge zone is extensively fractured and permeable but less than 100 ft thick. As the unit thickens toward the western end of the range, it is estimated that discharges may exceed 500 gal/min, approaching discharges reported in the Wind River Mountains. At the southeastern end of the Wind River Mountains, a flowing well (SB30–96–07b) discharges from the Park City Formation at a rate of 700 gal/min (Whitcomb and Lowry, 1968). At the northern end of the Wind River Mountains, Kendall Cold Spring (SB38–110–25) discharges from the Phosphoria Formation at a rate of 628 gal/min (Rouse, 1967, p. 3–4). As noted previously, Kendall Warm Spring, with a discharge of about 2,900 gal/min, also discharges from the Phosphoria Formation in the Wind River Mountains. However, as shown in figure 88, the discharge from this spring clearly is an outlier for the Park City–State Bridge zone. For reasons explained previously, Kendall Warm Spring probably is an outlet for water rising from the Tensleep Sandstone along the Wind River fault.

GROUND-WATER MOVEMENT

Though aquifers and confining units can be distinguished by hydrologic properties, it is apparent from the preceding discussion that all hydrogeologic units composed of Paleozoic rocks in the UCRB store and transmit water. Ground water moves not only from recharge areas to discharge areas within aquifers but from aquifers to aquifers through confining units throughout the UCRB. Primarily, ground water flows from structurally and topographically raised areas to structural basins and the incised canyons of the Colorado and Green Rivers and major tributaries. Movement occurs along regional, intermediate (subregional), and local flow paths (see Freeze and Cherry, 1979, p. 192–199, for a discussion of flow paths).

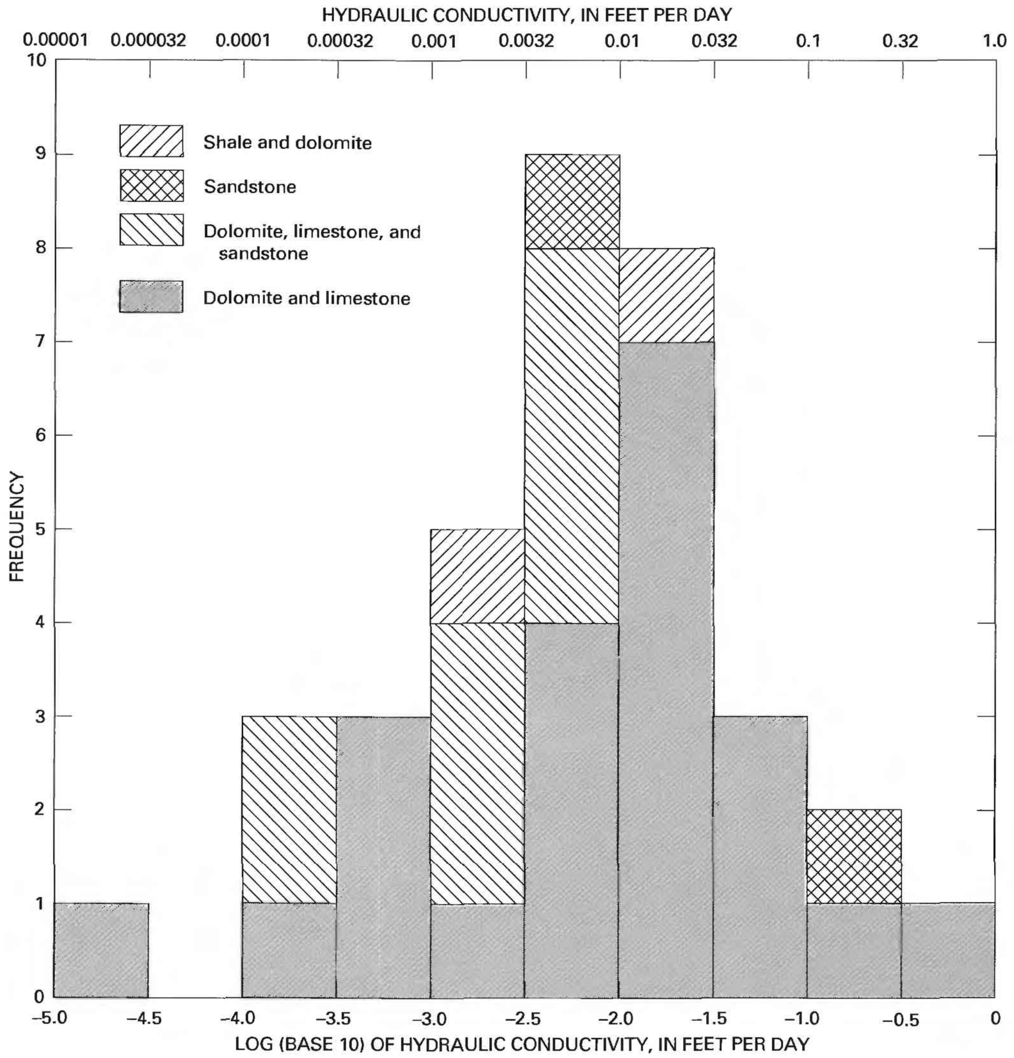


FIGURE 87.—Frequency distribution of hydraulic conductivity in the Park City–State Bridge zone of the Canyonlands aquifer.

Within the Four Corners aquifer system, aquifers and confining units are connected hydraulically by fractures, solution channels, and permeable beds, and head differences from the top to the bottom of the aquifer system generally are less than 500 ft. The Redwall-Leadville zone of the Madison aquifer transmits most of the water in the Four Corners aquifer system, and heads within this hydrogeologic unit define the potentiometric surface for the aquifer system (pl. 11). Subregional aquifers and confining units in the Four Corners aquifer system transmit water as lithologic, structural, topographic, and weathering conditions permit.

The Four Corners confining unit separates the Four Corners aquifer system from the Canyonlands aquifer, except in areas of structural disturbance or where the confining unit is missing because of erosion or nondeposition. Because of the effectiveness of the Four Corners confining unit in retarding vertical groundwater movement, head differences between the Canyonlands aquifer and Four Corners aquifer system in many areas exceed 1,000 ft (fig. 89).

In the Canyonlands aquifer, the Weber-De Chelly zone transmits most of the water. Where the Weber-De Chelly zone pinches out and becomes unsaturated in the eastern part of the UCRB, the Cutler-Maroon zone becomes the dominant water-transmitting zone within the Canyonlands aquifer. The potentiometric surface of the Canyonlands aquifer (pl. 11) is a composite of heads in the Weber-De Chelly and Cutler-Maroon zones where each zone is dominant. In the Cutler-Maroon and Park City–State Bridge zones, water is transmitted as lithologic, structural, topographic, or weathering conditions permit. Contiguous permeable intervals in some areas hydraulically connect the zones within the Canyonlands aquifer; in other areas, thick intervals of negligibly permeable rocks isolate permeable intervals within zones or sever hydraulic connection between zones within the aquifer.

Vertical head gradients between the Canyonlands aquifer and the Four Corners aquifer system indicate that water can be recharged to the Four Corners aquifer system by downward movement from the Canyonlands aquifer even in areas where the Four Corners aquifer system is deeply buried. In other areas, vertical head gradients favor upward movement of water from the Four Corners aquifer system to the Canyonlands aquifer. In figure 89, for example, vertical head gradients between the Canyonlands aquifer and Four Corners aquifer system indicate a potential for downward ground-water movement in the Monument Upwarp and upward ground-water movement in the Blanding Basin.

Patterns of ground-water movement within the Paleozoic rocks are complicated further by lithologic variability within underlying Precambrian rocks and overlying Mesozoic rocks. As a result of this variability, distinctions in some areas between hydrogeologic units composed of Paleozoic rocks and those composed of Precambrian or Mesozoic rocks are somewhat arbitrary. For example, in the Uinta Mountains and vicinity, in the Lees Ferry area, and in a few other areas throughout the UCRB where the uppermost Precambrian rocks are sedimentary (fig. 15), Precambrian sandstone, conglomerate, and limestone probably transmit water together with the Four Corners aquifer system. In areas where the lowermost Mesozoic rocks consist largely of sandstone, conglomerate, or limestone (fig. 17), the Mesozoic rocks commonly are connected hydraulically to the Canyonlands aquifer. For example, in the Defiance Plateau area, many wells draw water from the De Chelly Sandstone of Permian age and the overlying Shinarump Conglomerate Member of the Chinle Formation of Triassic age (Davis and others, 1963). In some areas where the uppermost or lowermost Paleozoic formations consist mostly of shale or quartzite, the Paleozoic rocks have more in common with the underlying basal and overlying Chinle-Moenkopi confining units than with the hydrogeologic units to which the Paleozoic rocks have been assigned. For example, where the State Bridge Formation, Goose Egg Formation, or shaly members of the Park City and Phosphoria Formations comprise the uppermost Paleozoic rocks, the uppermost Paleozoic rocks and lowermost Triassic rocks generally function together to impede ground-water movement.

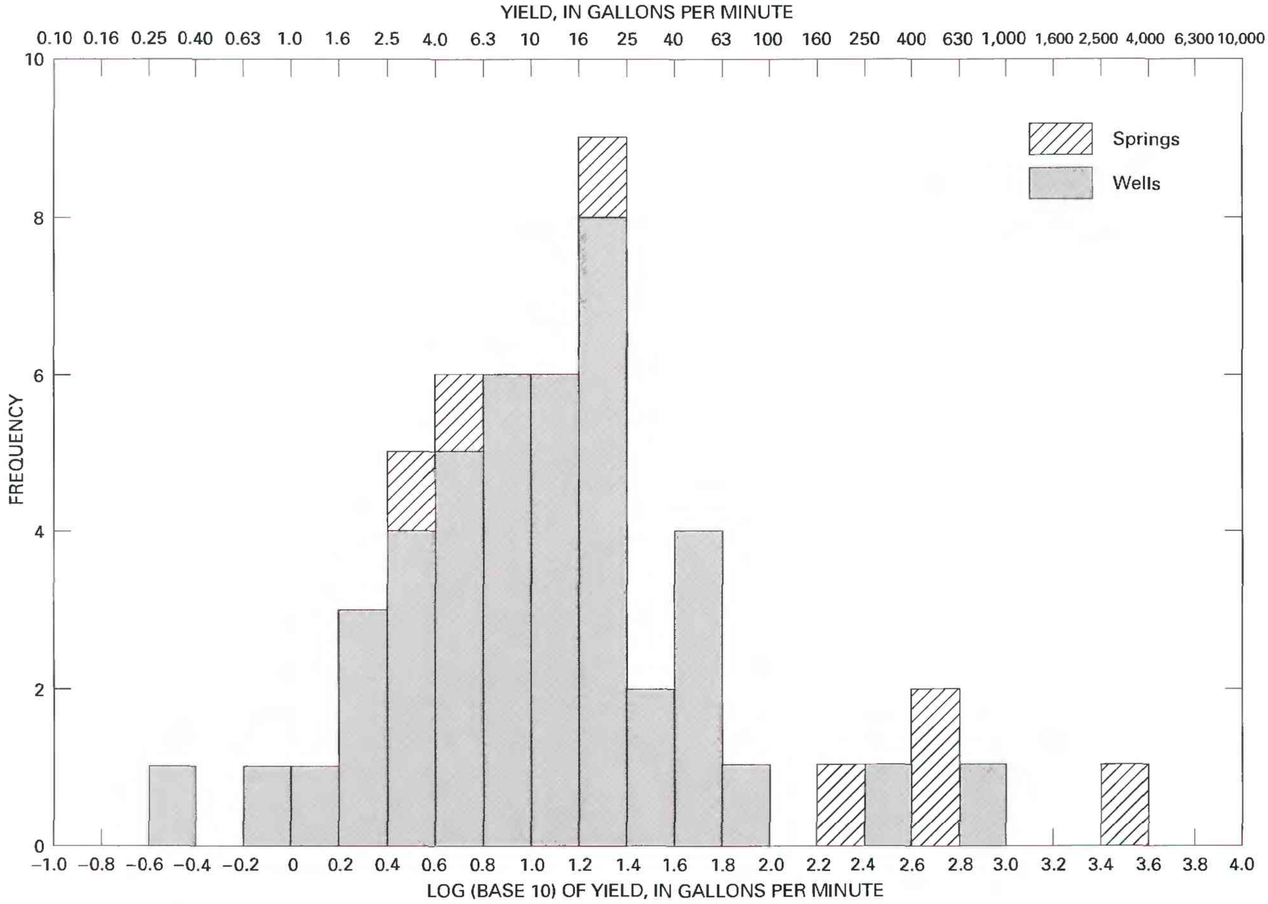


FIGURE 88.—Frequency distribution of yields from the Park City–State Bridge zone of the Canyonlands aquifer.

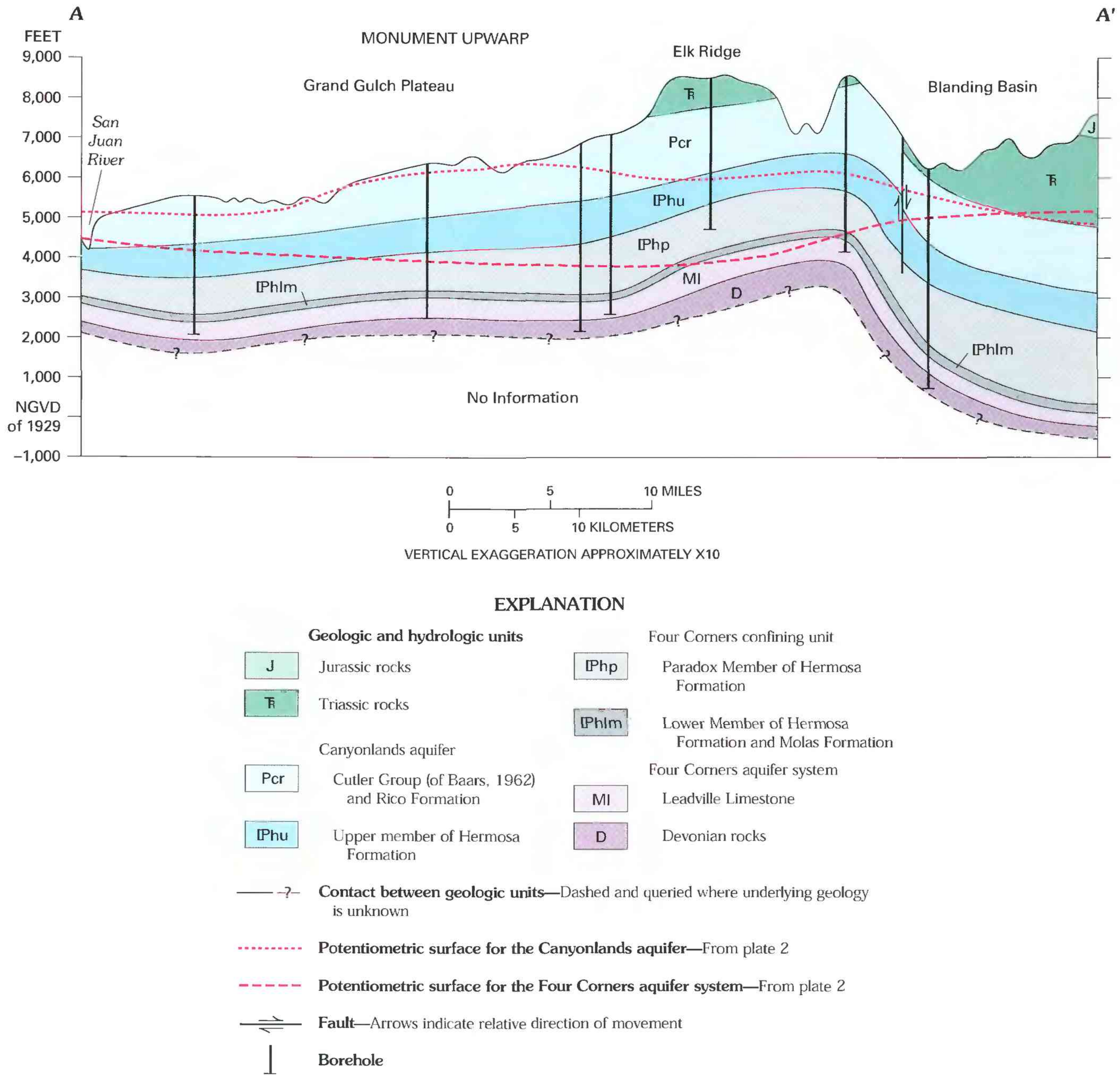


FIGURE 89.—Geologic section showing head differences between the Canyonlands aquifer and the Madison aquifer of the Four Corners aquifer system in southeastern Utah. (Modified from Thackston and others, 1981, p. 210. Location of section shown in fig. 14.)

RECHARGE TO THE PALEOZOIC ROCKS

Recharge to the Paleozoic rocks occurs from precipitation, either directly or by streamflow losses, and from interbasin flow. In general, recharge areas are characterized by downward ground-water movement, losing streams, unsaturated rock, water-table conditions, and fresh ground water. In contrast, discharge areas are characterized by predominantly upward ground-water movement, gaining streams, springs, flowing artesian wells, and brackish to briny ground water.

The largest potential for recharge exists where head differences between the Canyonlands aquifer and the Four Corners aquifer system are the most positive. Head differences between the Canyonlands aquifer and the Redwall-Leadville zone of the

Madison aquifer, as indicated on plate 12, range from less than –3,500 to more than +3,500 ft. A large potential for recharge can be inferred where head differences are more positive than +3,500 ft. Among areas where such large head differences occur are the Wind River, Uinta, Elk, Abajo, and Elkhead Mountains; La Sal Mountains; the Sierra Madre; the Park and Sawatch Ranges; the Uncompahgre and Kaibito Plateaus; the Cattle Creek Plateau (southern) section of the White River Plateau; the Wasatch and Fish Lake Plateaus (northern) section of the High Plateaus region; the Tavaputs Plateau, Roan Plateau, and Grand Mesa sections of the Uinta and Piceance Basins; the Sweetwater Arch; an arcuate area extending from the San Juan Mountains through the Sage Plain and Elk Ridge to the Grand Gulch Plateau; and several grabens in the Paradox Basin, including the Verdure, Lisbon Valley, and Big Gypsum Creek grabens. According to Metzger (1961) and Cooley and others (1969), predominantly downward ground-water movement also prevails in the Kaibab Plateau area.

Recharge and discharge areas and ground-water flow paths can be discerned from hydrochemical data as well as potentiometric heads. Areas in which the Redwall-Leadville zone of the Madison aquifer and the Weber-De Chelly zone of the Canyonlands aquifer contain freshwater (dissolved-solids concentrations of 1,000 mg/L or less) can be regarded as recharge areas for the Paleozoic rocks. Areas in which the Redwall-Leadville and Weber-De Chelly zones contain brackish to saline or briny water (dissolved-solids concentrations of 1,000 to as much as 300,000 mg/L) can be regarded as discharge areas for the Paleozoic rocks. The distribution of dissolved solids in water from the Redwall-Leadville and Weber-De Chelly zones is shown in figures 90 and 91. Together with topography, the distribution of average annual precipitation, the locations of losing and gaining reaches of streams, and the locations of springs, hydrochemical and potentiometric head data indicate that numerous recharge areas for the Paleozoic rocks exist within and peripheral to the UCRB (fig. 92).

PRECIPITATION

Precipitation is the major source of recharge to the Paleozoic rocks. Rain and snow falling on mountaintops, plateaus, ridges, mesas, and buttes infiltrate outcropping Paleozoic rocks or slowly percolate down through overlying volcanic and sedimentary rocks. Snowcover and storms of several days' duration prolong recharge. In contrast, summer thunderstorms provide little recharge because most of the rain from these storms immediately evaporates or runs off into stream channels. Recharge from precipitation is possible in areas receiving as little as 10 inches of average annual precipitation. For example, the Rawlins Uplift, with 11 inches of average annual precipitation, the San Rafael Swell, with 8 to 12 inches of average annual precipitation, and the Rainbow Plateau, with about 10 to 12 inches of average annual precipitation, all are reported to be recharge areas for Paleozoic and Mesozoic rocks (Berry, 1960; Cooley and others, 1969; Hood and Patterson, 1984). In most areas, including structural basins, altitudes above 5,900 ft generally receive at least 10 inches of precipitation in most years (fig. 11). On the basis of climatic studies in the region (previously cited in this report), it is estimated that evapotranspiration consumes between 45 and 90 percent of precipitation at altitudes between 6,000 and 9,000 ft, less precipitation at higher altitudes, and nearly all precipitation at lower altitudes.

Because of the lack of data, it is impossible to accurately determine the average annual evapotranspiration and, hence, the average annual recharge from precipitation to ground-water systems in the UCRB, excluding the San Juan Basin. However, for the entire UCRB, Iorns and others (1965, p. 10) calculated that the average annual precipitation was about 93,000,000 acre-ft and the average annual evaporation from water surfaces was 575,000 acre-ft (0.6 percent of the average annual precipitation). Subtraction of the water-surface evaporation and known stream outflows from the UCRB from the average annual precipitation would result in an unrealistically large estimate of recharge.

In a study of the northern Uinta Basin, encompassing an altitude range of 4,650 to 13,000 ft (similar to the entire UCRB) and an area of 5,200 mi^2, Hood and Fields (1978, p. 20) estimated recharge as a function of topography and precipitation and concluded that only about 10 percent of the average annual precipitation was available for recharge to aquifers in Paleozoic, Mesozoic, and Tertiary rocks in the area (table 17).

The results of a ground-water modeling study in southeastern Utah by Dunbar and Thackston (1985) also are enlightening. This study area was approximately 6,140 mi^2 and encompassed an altitude range of 3,500 to 12,700 ft. Direct surface recharge to the uppermost layer of Paleozoic rocks was estimated as a function of precipitation rates, topography, and the extent of outcrops of the Triassic Moenkopi and Chinle Formations (which in the area of the study comprise the lowermost confining unit above the Paleozoic rocks—the Chinle-Moenkopi confining unit). Through calibration, empirical relations developed by Eakin and others (1951) and Fiero (1968) applied to the local topographic and precipitation data were modified, resulting in estimates of recharge ranging from 0.05 to 0.55 in/yr in areas where the Paleozoic rocks are not covered by the Chinle-Moenkopi confining unit. The total recharge to Paleozoic rocks in the area was estimated to be 8.09 ft^3/s (5,900 acre-ft/yr), which would indicate that 99.8 percent of the ambient average annual precipitation (3,100,000 acre-ft/yr, according to Dunbar and Thackston, 1985, fig. 5.1) does not reach the Paleozoic rocks in the area of the study.

Given the different methods of estimating recharge from precipitation in past studies within the UCRB, any figure presented in this report must be considered tentative. If one uses the simplified method of Hood and Fields (1978), then, as shown in table 18, the total precipitation influx to the UCRB, excluding the San Juan Basin, is estimated to be 83,000,000 acre-ft/yr and the recharge available to ground-water systems (not necessarily

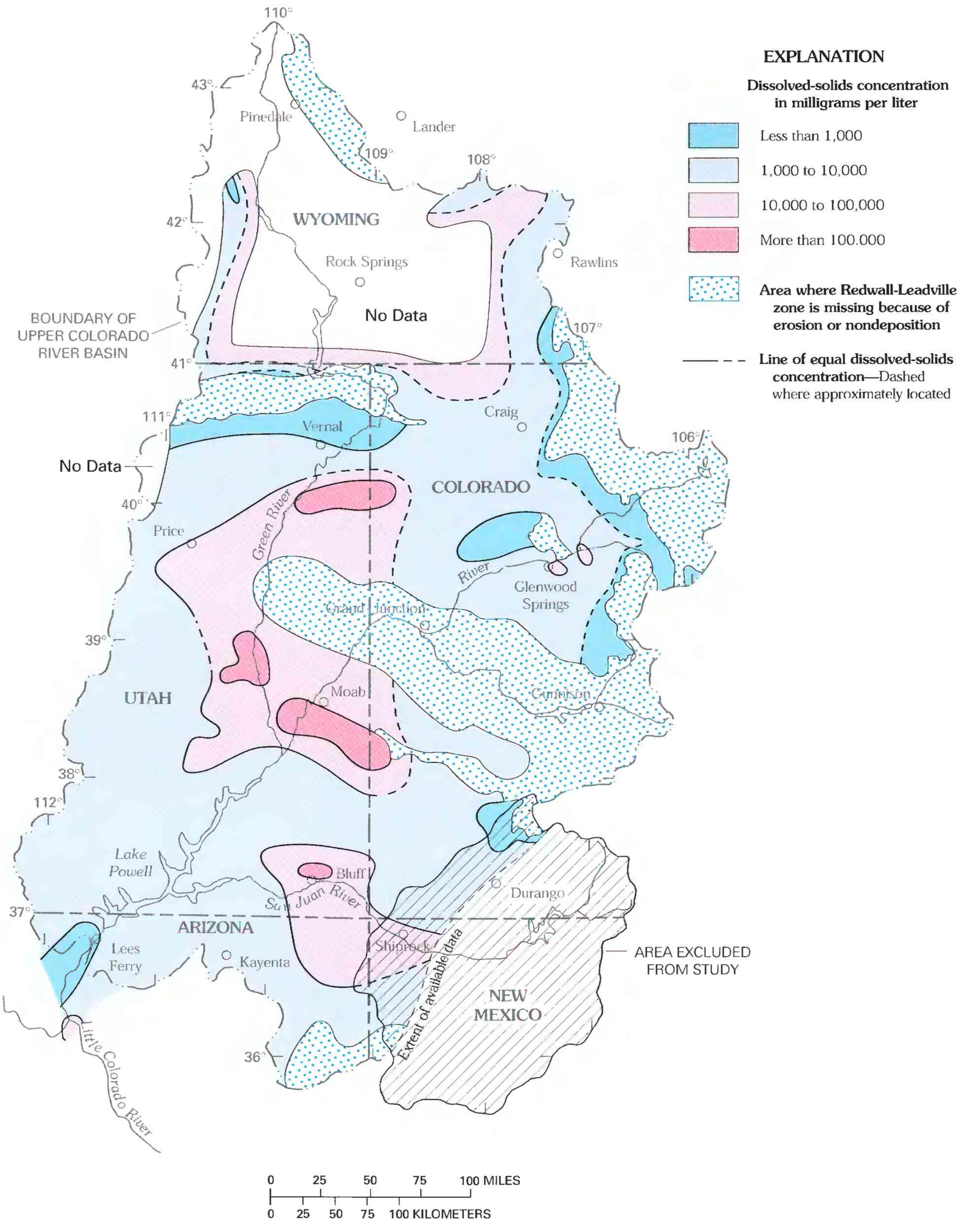


FIGURE 90.—Concentration of dissolved solids in water from the Redwall-Leadville zone of the Madison aquifer. (Modified from Lindner-Lunsford and others, 1985; Briant Kimball, U.S. Geological Survey, written commun., 1990.)

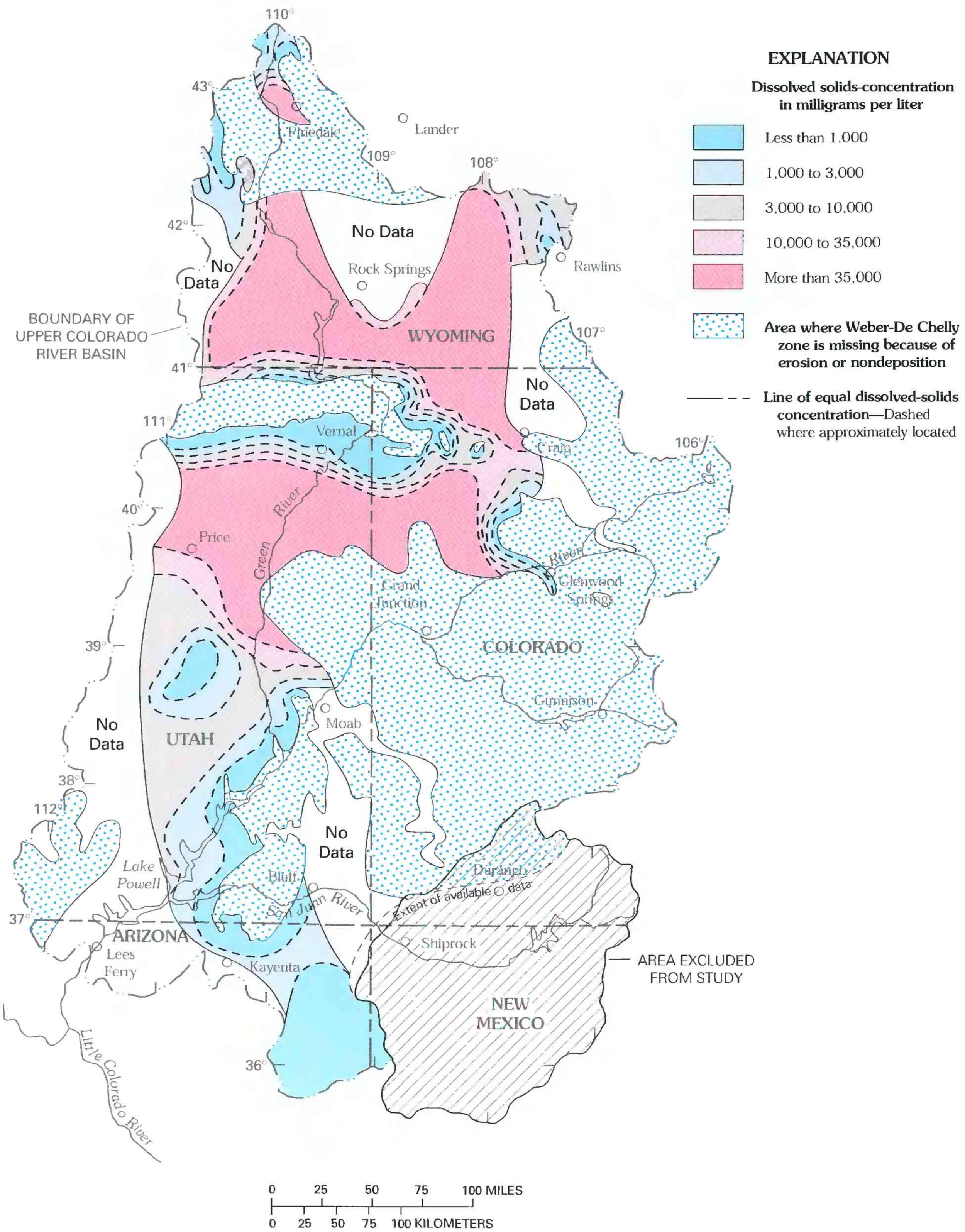


FIGURE 91.—Concentration of dissolved solids in water from the Weber-De Chelly zone of the Canyonlands aquifer. (Modified from Lindner-Lunsford and others, 1985.)

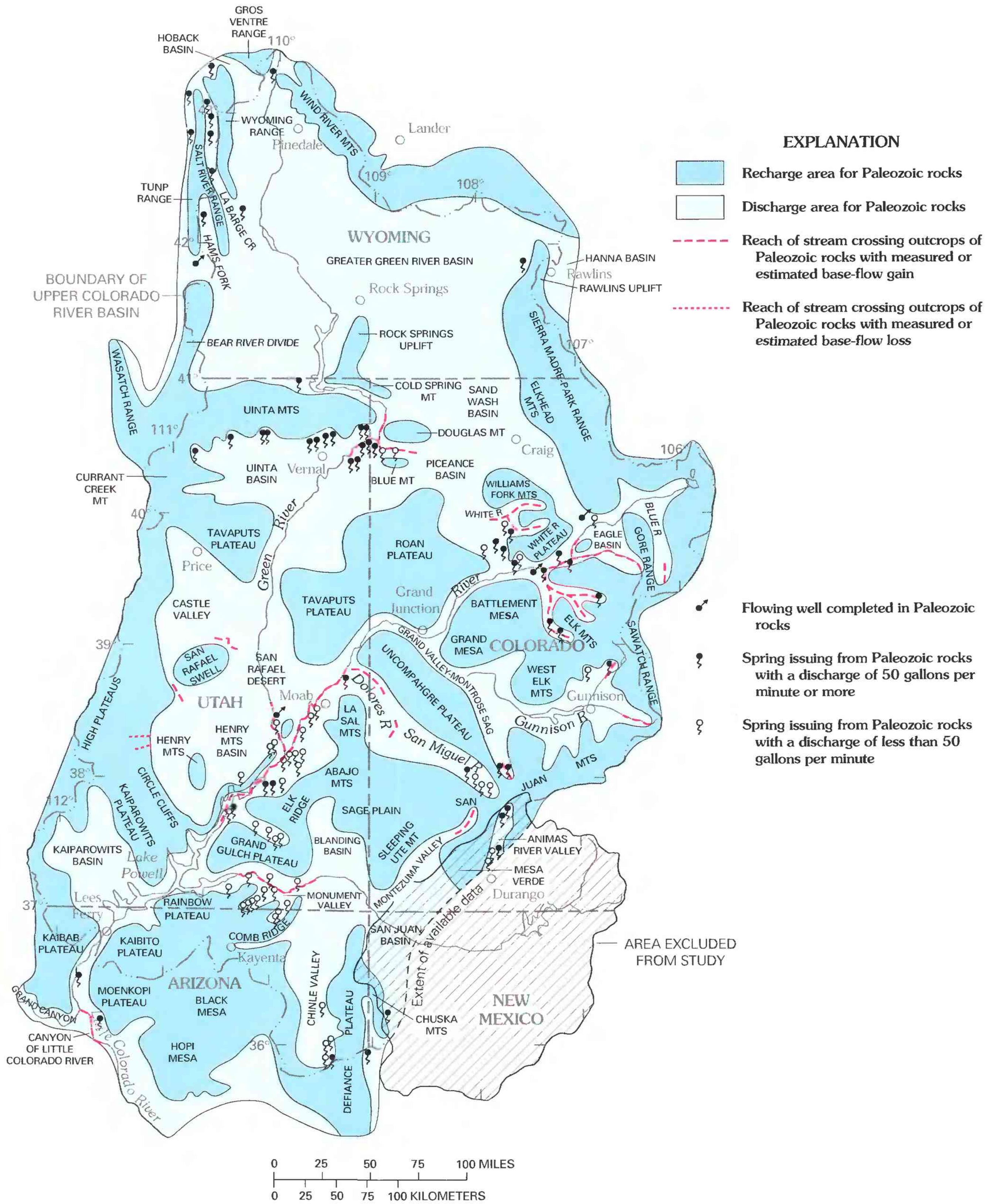


FIGURE 92.—Recharge and discharge areas for the Paleozoic rocks in the Upper Colorado River Basin.

TABLE 17.—*Estimated average annual volumes of precipitation and ground-water recharge, 1941–70, in the northern Uinta Basin and Uinta Mountains (Hood and Fields, 1978)*

[less than; ≤ , equal to or less than]

Precipitation zone (inches)	Acres (acres)	Estimated average annual precipitation		Estimated average annual ground-water recharge	
		Feet	Acre-feet	Percentage of precipitation	Acre-feet
<8	295,800	0.58	171,600	0	0
8–10	507,200	.75	380,400	0	0
10–12	293,000	.92	269,600	1	2,700
12–14	258,000	1.08	278,600	2	5,600
14–16	241,500	1.25	301,900	2	6,000
16–18	241,600	1.41	340,700	5	17,000
18–20	230,800	1.57	362,400	5	18,100
20–22	221,300	1.73	382,800	10	38,300
22–24	165,900	1.89	313,600	10	31,400
24–26	132,100	2.06	272,100	15	40,800
26–30	243,700	2.31	562,900	15	84,400
30–34	266,000	2.64	702,200	20	140,400
34–38	167,300	2.96	495,200	25	123,800
38≤42	12,500	3.25	40,600	25	10,200
Totals (rounded)	3,277,000		4,870,000		500,000

TABLE 18.—*Estimated average annual volumes of precipitation and ground-water recharge, 1931–80, in the Upper Colorado River Basin, excluding the San Juan Basin (based on planimeter measurements of contoured areas in fig. 10)*

[<, less than; >, greater than]

Precipitation zone (inches)	Area (acres)	Estimated average annual precipitation		Estimated average annual ground-water recharge	
		Feet	Acre-feet	Percentage of precipitation	Acre-feet
<6	2,209,939	0.42	928,174	0	0
6–10	21,344,557	.67	14,300,853	0	0
10–20	29,765,971	1.25	37,207,464	3	1,116,224
20–30	7,912,128	2.08	16,457,226	13	2,139,439
30–40	3,919,686	2.92	11,445,483	23	2,632,460
>40	745,741	3.75	2,796,529	25	699,132
Totals (rounded)	66,000,000		83,000,000		6,600,000

to aquifers composed of Paleozoic rocks) is 6,600,000 acre-ft/yr (1.2 in/yr). If Dunbar and Thackston (1985) are correct, most of this potential recharge is captured by aquifers in Mesozoic and Tertiary rocks, and little of this water moves downward into the Paleozoic rocks.

INTERBASIN FLOW

As seen on plate 11, surface-water and ground-water divides peripheral to the UCRB coincide in all but a few areas. However, potentiometric contours indicate that ground water flows into the UCRB from the Sierra Madre, the south flank of the Sweetwater Arch, the Hoback Basin, the Overthrust Belt, the east flank of the Wasatch Range and northwestern corner of the Uinta Basin, the High Plateaus region, and the Kaibab Plateau. In each of these areas, Quaternary surface-water divides have been superimposed upon preexisting structures. The Paleozoic rocks of the UCRB extend beneath these surface-water divides and crop out in the areas providing recharge to the UCRB. Distances to these recharge areas from the boundary of the UCRB range from less than 1 to 21 mi.

The volume of interbasin flow providing recharge to Paleozoic rocks in the UCRB can be estimated from Darcy's Law (Freeze and Cherry, 1979, p. 16) using topographic maps, geologic maps (figs. 14, 41, 65, and 74), and hydraulic gradients and transmissivity values for the Madison and Canyonlands aquifers (from pls. 3, 6, 8, and 11). This approach assumes that flow into the UCRB from hydrogeologic units other than the Redwall-Leadville zone of the Madison aquifer and the Cutler-Maroon and Weber-De Chelly zones of the Canyonlands aquifer is inconsequential. As shown in table 19, the total volume of interbasin flow contributing recharge to Paleozoic rocks in the UCRB is estimated to be about 1,000 acre-ft/yr. The recharge from interbasin flow is less than 1 percent of the estimated recharge from precipitation available to aquifers in Pale'nsozoic, Mesozoic, and Tertiary rocks in the UCRB.

STREAMFLOW

Not all of the precipitation that becomes recharge directly infiltrates aquifers in the UCRB. Some of this precipitation runs off into streams and then infiltrates bedrock beneath or along the channels. Ephemeral streams crossing outcropping or shallowly buried Paleozoic rocks are believed to provide recharge in the Rawlins Uplift (Berry, 1960, p. 31), Sierra Madre (Welder and McGreevy, 1966), San Rafael Swell (Hood and Patterson, 1984, p. 30), Dirty Devil River Basin (Hood and Danielson, 1981, p. 32), northwestern Paradox Basin (Rush and others, 1982, p. 29), Navajo and Hopi Indian Reservations (Cooley and others, 1969, p. 40), and northern Uinta Basin (Hood and Fields, 1978, p. 32–33). In the latter area, recharge from streams is estimated to be only 4 percent of the average annual recharge to the aquifers that are present.

Streamflow losses from perennial streams to Paleozoic rocks can be substantiated in only a few areas. Where the Dolores River crosses the Big Gypsum graben, the base-flow discharge of the river decreases from 38 to 31 ft^3/s (Warner and others, 1985). Most of the lost streamflow is believed to seep into Mesozoic rocks along the channel, but some may be infiltrating outcropping carbonate, clastic, and evaporitic rocks of the Pennsylvanian Hermosa Formation. In the headwater reaches of Ashley Creek, Brush Creek, and Little Brush Creek, all or part of the streamflow in individual creeks disappears into Mississippian carbonate rocks through in-channel sinkholes and caves (Maxwell and others, 1971). Most of the lost streamflow emerges shortly downstream as springs issuing from Pennsylvanian, Permian, and Triassic formations. Some, however, remains as ground water in the Paleozoic rocks (Feltis, 1966, p. 12).

FLOW PATHS

Because of highly variable topography within and peripheral to the UCRB, water in the Paleozoic rocks of the UCRB is forced to flow toward local and subregional outlets, rather than toward a single, regional discharge area (pl. 11). Although some ground water flows vertically through the Four Corners confining unit, the generally small permeability of this confining unit inhibits interchange of water between the underlying Four Corners aquifer system and the overlying Canyonlands aquifer. Ground-water movement in the Paleozoic rocks ultimately is directed mainly toward four areas—the eastern Great Divide Basin (between the Rawlins Uplift and Sierra Madre), the confluence of the Yampa and Green Rivers, the San Juan Basin, and the confluence of the Colorado and Little Colorado Rivers. The fourth area is in the Lower Colorado River Basin, just south of the politically defined boundaries of the UCRB.

CIRCULATION IN THE FOUR CORNERS AQUIFER SYSTEM

Ground-water circulation in the Four Corners aquifer system is constrained by the Continental Divide from the San Juan Mountains to the Sierra Madre and by potentiometric divides elsewhere. In this report, these potentiometric divides are called the Wind River–Northern Great Plains divide from the Sierra Madre through the Rawlins Uplift and Sweetwater Arch to the Wind River Mountains; the Upper Colorado River Basin–Basin and Range divide from the Gros Ventre Range and Overthrust Belt through the Uinta Mountains and Wasatch Range to the High Plateaus and Kaibab Plateau; the Black Mesa divide from the Kaibito Plateau through Black Mesa to the Defiance Plateau; and the San Juan–Chuska divide from the Chuska Mountains to the San Juan Mountains. Within the UCRB, subregional ground-water flow paths in the Four Corners aquifer system are influenced by six potentiometric divides—the Uinta-Park, Tavaputs, Grand Mesa–White River, Eagle, Circle Cliffs, and Monument divides (pl. 11).

TABLE 19.—*Estimated annual interbasin flow contributing recharge to Paleozoic rocks in the Upper Colorado River Basin*

[A question mark indicates the value is a rough estimate based on topographic and geologic maps and extrapolation of data from other areas]

Aquifer zones	Average transmissivity (feet squared per day)	Average hydraulic gradient	Flow-path width (feet)	Annual interbasin flow contributing recharge (acre-feet per year)
		Sierra Madre		
Weber-De Chelly zone	5.1	0.011	105,600	50
Cutler-Maroon zone		Missing		0
Redwall-Leadville zone		Ground-water and surface-water divides coincide		0
		Sweetwater Arch		
Weber-De Chelly zone	1.5	.052	279,840	183
Cutler-Maroon zone	.36 (?)	.052	279,840	44
Redwall-Leadville zone	.39 (?)	.019	369,600	23
		Hoback Basin		
Weber-De Chelly zone	1.7	.0086	121,440	15
Cutler-Maroon zone	.76 (?)	.0086	121,440	6.6
Redwall-Leadville zone	.81 (?)	.025	168,960	29
		Overthrust Belt		
Weber-De Chelly zone	8.5	.043	42,240	129
Cutler-Maroon zone	1.4 (?)	.043	42,240	21
Redwall-Leadville zone	1.2 (?)	.0094	15,840	1.5
		Uinta Basin		
Weber-De Chelly zone	5.6	.014	142,560	94
Cutler-Maroon zone		Missing		0
Redwall-Leadville zone	1.2 (?)	.018	279,840	50
		High Plateaus		
Weber-De Chelly zone	.64	.026	667,920	94
Cutler-Maroon zone	.90 (?)	.026	667,920	131
Redwall-Leadville zone	1.4 (?)	.029	401,280	138
		Kaibab Plateau		
Weber-De Chelly zone		Missing		0
Cutler-Maroon zone	1.4 (?)	.047	174,240	96
Redwall-Leadville zone		Ground-water and surface-water divides coincide		0
Total				1,120

North of the Uinta-Park divide, water in the Four Corners aquifer system primarily moves from peripheral and internal highlands toward the eastern edge of the Great Divide Basin, where it flows through a gap between the Rawlins Uplift and Sierra Madre into the Hanna Basin. Large springs, such as Hogsback Spring (SB26–114–01bac) and Sheep Creek Spring (SLA02–19–16bcb) in the Overthrust Belt, oil and gas field pumpage in the Overthrust Belt and Rock Springs Uplift (table 20), and a few water wells in the Rawlins area deplete some of the water in circulation. In the interior of the greater Green River and Sand Wash Basins, water under the influence of strongly upward hydraulic gradients seeps up into Pennsylvanian and Permian rocks.

Between the Uinta-Park and Tavaputs divides, circulation in the Four Corners aquifer system is directed primarily toward the confluence of the Yampa and Green Rivers on the southeast flank of the Uinta Mountains (Geldon, 1986). However, some of the water in circulation issues as springs in recharge areas, and some flow is directed toward the Glenwood Springs area and Eagle Basin by the Grand Mesa–White River and Eagle divides.

Considerable volumes of water discharge from the aquifer system not far from the recharge areas between the Uinta-Park and Tavaputs divides. For example, Teller and Welder (1983, p. 12) indicate that springs and seeps entering a 4-mi reach of Rifle Creek where it is incised into the White River Plateau upstream from the Rifle Falls Fish Hatchery have a combined discharge of 28 ft^3/s. Bowles Fish Hatchery Spring (SC08–83–09), in the Sawatch Range, has a discharge of 1,123 gal/min (Iorns and others, 1964).

According to URS Corp. (1983), the combined discharge of seeps and springs into Glenwood Canyon between Dotsero and Glenwood Springs is 30.2 ft^3/s; the combined flow of hot springs and seeps at Glenwood Springs is about 4,300 gal/min (Geldon, 1989c). The topography and geology of the

TABLE 20.—*December 1986 water production from oil and gas fields where the principal reservoir is in Paleozoic rocks of the Upper Colorado River Basin*

[Location of most oil and gas fields shown on pl. 11. Source of data is Petroleum Information Corp., written commun., 1987]

Field	State	Type	Principal Paleozoic reservoirs(s)	Other reservoirs	Water production	
					Gallons per minute	Cubic feet per second
			Canyonlands aquifer			
Lost Soldier Wertz-Mahoney	Wyoming	Oil, gas	Tensleep Sandstone, Amsden Formation	Madison Limestone, Flathead Sandstone, Triassic, Jurassic, and Cretaceous rocks	8,620	19.2
Brady	Wyoming	Oil, gas	Weber Sandstone	Jurassic and Cretaceous rocks	311	.69
O'Brien Springs	Wyoming	Oil	Tensleep Sandstone	Jurassic and Cretaceous rocks	109	.24
Happy Springs	Wyoming	Oil	Park City Formation	Cretaceous rocks	6.9	.015
Butcher Knife Springs	Wyoming	Gas	Morgan Formation	Cretaceous rocks	.23	.00051
Moxa	Wyoming	Gas	Morgan Formation	Cretaceous rocks	.058	.00013
Rangely	Colorado	Oil	Weber Sandstone	Triassic, Jurassic, and Cretaceous rocks	15,046	33.5
Ashley Valley	Utah	Oil	Weber Sandstone, Park City Formation	Jurassic and Cretaceous rocks	1,086	2.4
Elk Springs	Colorado	Oil	Weber Sandstone		17	.038
Moffat	Colorado	Oil	Weber Sandstone and Pennsylvanian rocks	Triassic, Jurassic, and Cretaceous rocks	3.5	.0078
Danforth Hills	Colorado	Oil	Weber Sandstone	Triassic, Jurassic, and Cretaceous rocks	.12	.00027
Upper Valley	Utah	Oil	Permian rocks		578	1.3
Boundary Butte	Utah, Arizona	Oil, gas	Cutler and Hermosa Formations		65	.15
Barker Dome	Colorado	Oil	Hermosa Formation		1.6	.0036
Ferron	Utah	Oil	Permian rocks		.12	.00027
Subtotal					25,845	57.6
			Four Corners confining unit			
Greater Aneth	Utah, Colorado	Oil	Paradox Member of Hermosa Formation		2,243	5.0
Marble Wash-Towaoc	Colorado	Oil	Paradox Member of Hermosa Formation		1.2	.0026
Long Canyon	Utah	Oil	Paradox Member of Hermosa Formation		.45	.0019
Bug	Utah	Oil	Paradox Member of Hermosa Formation		.28	.062
Tohonadla	Utah	Oil	Paradox Member of Hermosa Formation		.96	.0021
Patterson Canyon	Utah	Oil	Paradox Member of Hermosa Formation		4.8	.011
Turner Bluff-Cowboy	Utah	Oil	Paradox Member of Hermosa Formation		3.2	.0071
Papoose Canyon	Colorado	Oil	Paradox Member of Hermosa Formation		.56	.0012
Teec Nos Pos	Arizona	Gas	Paradox Member of Hermosa Formation		.44	.00098
Subtotal					2,255	5.09
			Four Corners aquifer system			
Table Rock	Wyoming	Gas	Madison Limestone	Weber Sandstone, Jurassic, Cretaceous, and Tertiary rocks	45	0.10
Lake Ridge-Fogarty Creek-Dry Piney	Wyoming	Gas	Madison Limestone, Bighorn Dolomite	Jurassic and Cretaceous rocks	121	.27
Greater Lisbon	Utah	Oil, Gas	Leadville Limestone, Devonian rocks	Permian rocks	164	.37
Salt Wash	Utah	Oil	Redwall Limestone		.56	.0012
Dry Mesa	Arizona	Oil	Leadville Limestone		3.7	.0082
McElmo	Colorado	Oil	Leadville Limestone	Cutler Formation	18	.040
Subtotal					352.3	0.79
Total					28,452	63.5

Glenwood Springs area and the chemistry of water discharging at Glenwood Springs indicate that most of the water reaching Glenwood Springs is from areas to the south (pl. 11 and fig. 93). The water discharging at Glenwood Springs is a sodium chloride type with an average dissolved-solids concentration of about 20,000 mg/L (Geldon, 1989c). Plots of log Q/K (defined in fig. 94) against temperature, drawn from output of the speciation program SOLVEQ (Spycher and Reed, 1989a) for minerals in equilibrium with water discharging from the Leadville Limestone at Glenwood Springs, indicate a water-rock equilibration temperature of about 90°C (fig. 94). Assuming this equilibration temperature and recharge similar in composition to the least evolved water from the Maroon Formation in the Crystal River-Cattle Creek area (Brogden and Giles, 1976a), the reaction-flow-path model CHILLER (Spycher and Reed, 1989b) successfully simulated the chemistry of the water discharging at Glenwood Springs (fig. 95) by the following flow path:

a. Meteoric water with a temperature of about 5°C infiltrates the Maroon Formation on the Cattle Creek Plateau or in the Elk Mountains.

b. Descending through about 9,000 ft of the Maroon Formation, Gothic Formation, Eagle Valley Evaporite, and Belden Formation, the water is heated to about 90°C and reacts sequentially with aluminosilicate minerals, gypsum, halite, and illite.

c. In the Leadville Limestone and Dyer Dolomite, the water travels toward Glenwood Springs, equilibrating with carbonate minerals along the flow path.

d. At Glenwood Springs, the water rises rapidly along faults, cools to 49°C, and in the process, precipitates micaceous minerals, kaolinite, calcite, hematite, and quartz.

The Continental Divide and the Eagle ground-water divide direct flow in the Four Corners aquifer system toward the center of the Eagle Basin where tremendous artesian pressures have

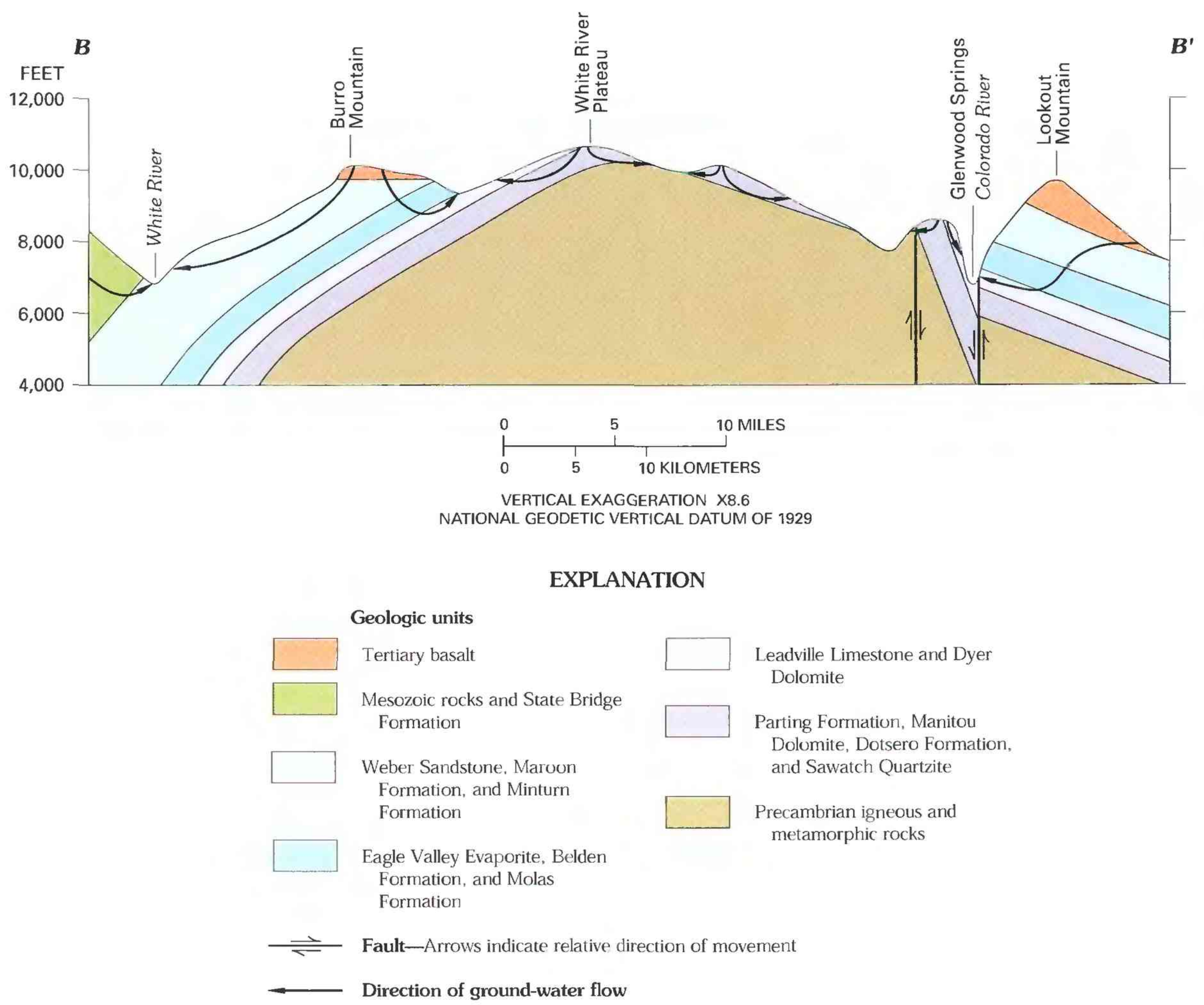


FIGURE 93.—Generalized geologic section across the White River Plateau showing ground-water circulation in the Paleozoic rocks (location of section is shown on pl. 11).

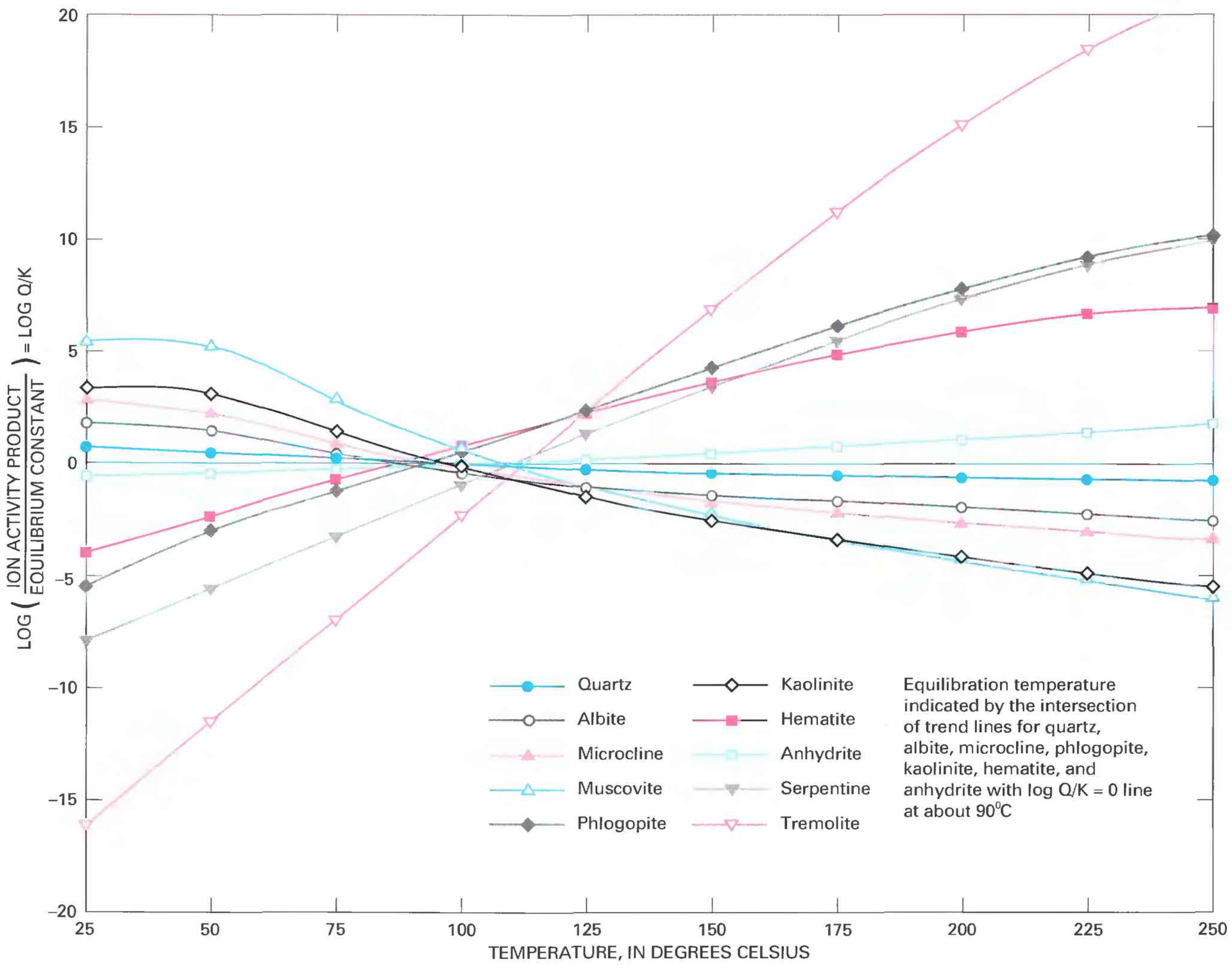


FIGURE 94.—Plot of log $\left(\frac{\text{Ion activity product}}{\text{Equilibrium constant}}\right)$ against temperature for water from the Leadville Limestone discharging from the Redstone 21–9 well at Glenwood Springs, Colorado.

built up. One well in this area (SC02–84–02bc), after penetrating the Leadville Limestone, created a geyser 300 ft high and produced a discharge that decreased with time from 9,500 to 3,200 gal/min (Hampton, 1974; Colorado Department of Natural Resources State Engineers Office, written commun., 1983). However, in 1988, this well and others in the area with similarly large discharges were not known to be in use.

Between the Tavaputs and Uinta-Park divides and west of the Grand Mesa–White River divide, several springs on the south flank of the Uinta Mountains, such as Big Spring (UB01–08–17cbb), Indian Big Spring (UB02–02–05dbb), and Pole Creek Spring (UB03–02–34d), discharge large volumes of water from the Four Corners aquifer system not far from sources of recharge. Pole Creek Spring, for example, has a discharge that varies seasonally from 2 to 25 ft^3/s (Maxwell and others, 1971, p. 15). However, considerable volumes of ground water do not discharge locally but flow to the far end of the subregional flow path and emerge at Split Mountain Warm Spring (SLD 04–24–16cdd), which has a discharge of 2,700 gal/min.

South of the Tavaputs divide and north of the Circle Cliffs and Monument divides, water in the Four Corners aquifer system moves west from the San Juan Mountains, south and east from the San Rafael Swell–Capitol Reef–High Plateaus area, and northwest from the Elk Ridge-Abajo Mountains area (Hanshaw and Hill, 1969, p. 271–272; INTERA Environmental Consultants, Inc., 1984, p. 59). Because of faulting and topography, several springs with an average combined discharge of 672 gal/min issue from the Leadville Limestone at Ouray, Colo. (David Vince, City of Ouray, written commun., 1988). The discharges of some of these springs are relatively constant, whereas others vary substantially throughout the year (fig. 96) in response to melting of the snowpack in the San Juan Mountains. Geothermal wells in use by the City of Ouray in 1988 flowed at a combined average rate of 388 gal/min. (Other geothermal wells have been completed at Ouray but were not in use in 1988.) In the western Paradox Basin and vicinity, ground water moves into closed depressions in the potentiometric surface, resulting in stagnation. According to Thackston and others (1981, p. 219), this stagnant water is trapped in density-driven convection cells. Some of the water in the Paradox Basin is pumped from oil and gas fields, such as the greater Lisbon oil field (table 20).

Between the Monument, Black Mesa, and San Juan–Chuska divides, all water in the Four Corners aquifer system apparently moves toward the vicinity of the San Juan River and the greater Aneth oil and gas fields (pl. 11). This water apparently moves up into the Four Corners confining unit, where it is pumped from oil and gas wells.

All water in the Four Corners aquifer system south of the Circle Cliffs divide and west of the Black Mesa and Monument divides that is not discharged along local flow paths to springs or pumped from wells ultimately discharges as springs into the canyon of the Little Colorado River or in Marble Canyon southwest of Lees Ferry, Ariz. (INTERA Environmental Consultants, Inc., 1984, p. 59). Some of the springs in this area are so low in dissolved solids (for example, Vasey's Paradise Spring, which has a dissolved-solids concentration of 163 mg/L) that they undoubtedly are sustained mostly by precipitation on plateaus bordering the canyons. Most of the springs, however, discharge water that has traveled farther from areas of recharge. Some of the water, in fact, may have traveled considerable distances through Pennsylvanian and Permian rocks and descended into the Four Corners aquifer system through fractures and faults associated with the East Kaibab monocline south and east of the Little

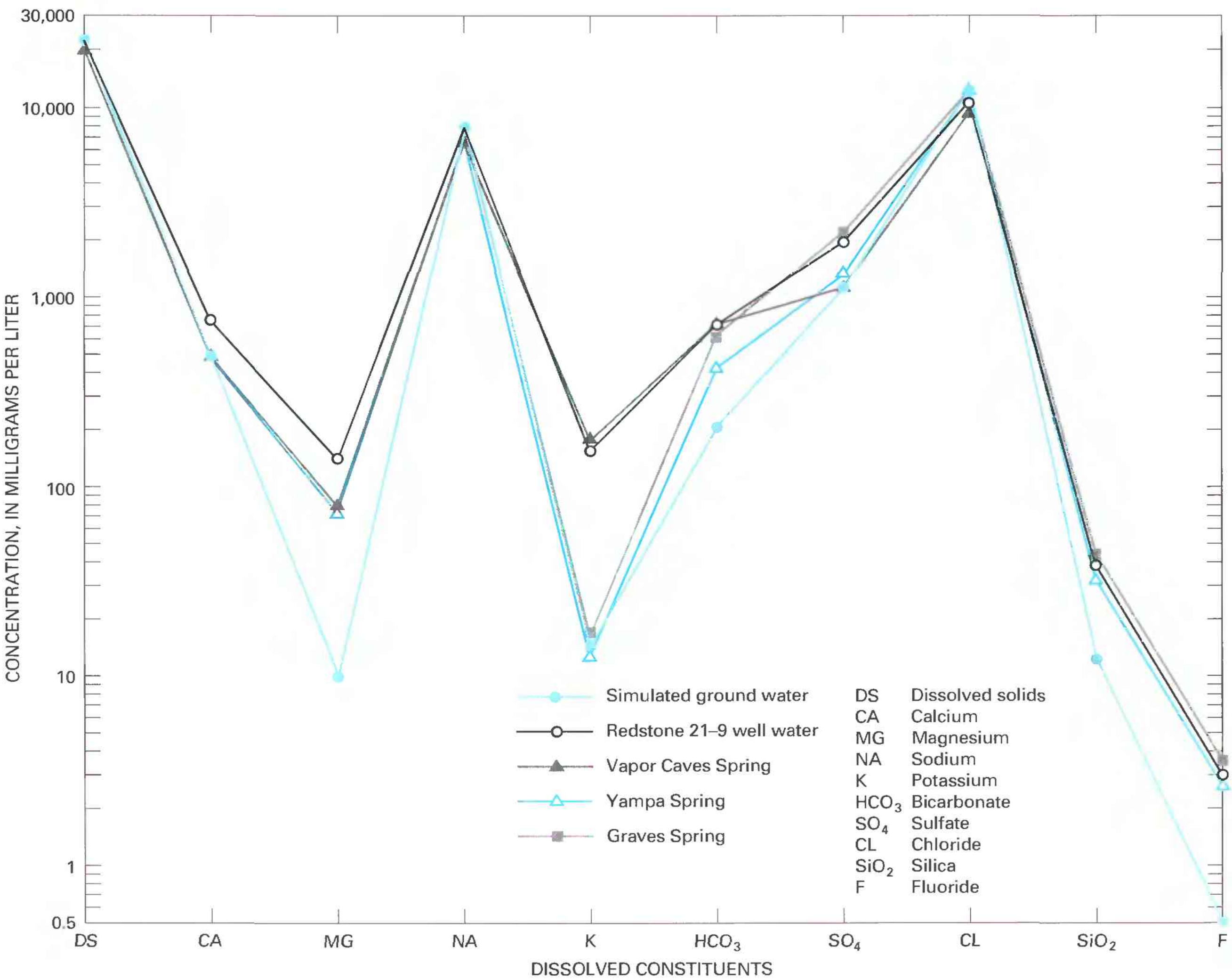


FIGURE 95.—Comparison of major ion chemistry modeled by CHILLER (Spycher and Reed, 1989b) and analyzed ground water from Leadville Limestone at Glenwood Springs, Colorado.

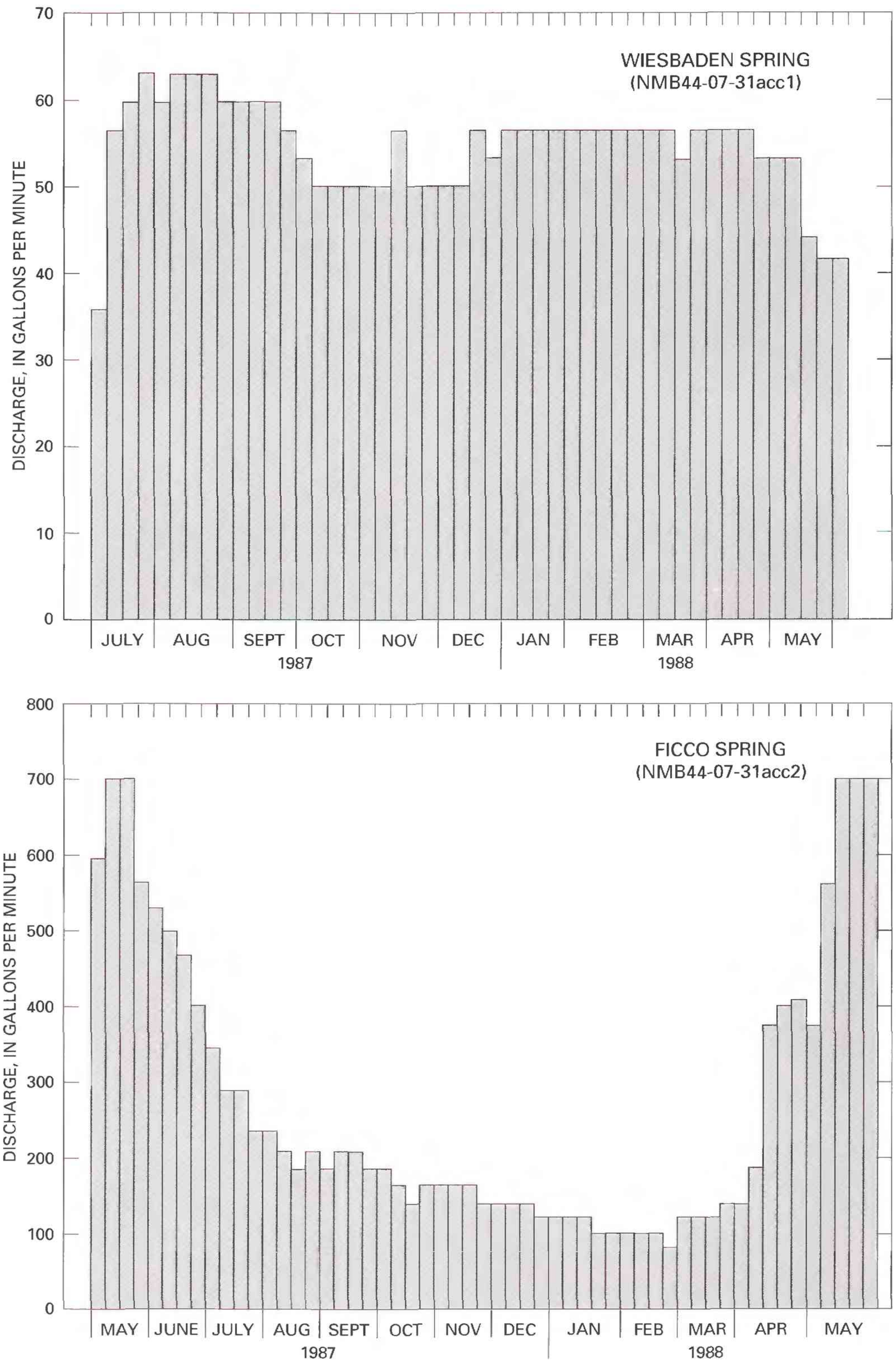


FIGURE 96.—Recorded discharges of thermal springs at Ouray, Colorado, 1987 to 1988 (provided by David Vince, City of Ouray, written commun., 1988).

Colorado River (Cooley, 1976, p. 8–9). The combined flow of all springs discharging in the canyon of the Little Colorado River averages about 100,000 gal/min (223 ft^3/s); Blue Spring, the largest single contributor, has a discharge of 42,000 to 45,000 gal/min (Cooley, 1976, p. 5–8). The combined flow of all springs discharging in Marble Canyon is about 5,000 gal/min (11 ft^3/s); Vasey's Paradise Spring (GA36–05–27b), the largest single contributor, has a discharge of 45 to 4,500 gal/min (Cooley, 1976; McCulley, 1985). An undetermined amount of the water discharging in the canyon of the Little Colorado River and in Marble Canyon originates in areas to the south of the UCRB and represents outflow from the Lower Colorado River Basin.

EFFECT OF THE FOUR CORNERS CONFINING UNIT ON CIRCULATION

Despite its effectiveness in inhibiting ground-water movement, the Four Corners confining unit contains water in isolated layers and lenses. According to Charles Spencer (U.S. Geological Survey, oral commun., 1986), much of this water is connate. Water pumped from the Paradox Member of the Hermosa Formation in some of the oil and gas fields listed in table 20 may fit this interpretation. Unfortunately, any interpretation of the effect of the Four Corners confining unit on ground-water movement is hindered by a paucity of data. Sufficient data for interpretation exist only for the Paradox Basin and adjacent areas of southeastern Utah and in the Eagle Basin and adjacent areas of northwestern Colorado.

In the Paradox Basin and adjacent areas, including the Henry Mountains Basin, Monument Upwarp, and Blanding Basin, the prevailing direction of ground-water movement through the Four Corners confining unit appears to be vertical. Seemingly in contradiction, Hanshaw and Hill (1969, p. 277) presented a map of the potentiometric surface for the Paradox Member of the Hermosa Formation indicating a regionally interconnected flow system. This map, however, is based primarily on head data for the marine shelf facies of the Paradox Member, a sequence of interlayered carbonate rocks and sandstone at the top and margins of the member. The bulk of the Paradox Member consists of shale and evaporites, not carbonate rocks and sandstone. When head data are plotted for the entire Paradox Member, as was done by Thackston and others (1981, p. 214), there appear to be no geographic trends. Thus, a regionally interconnected flow system is not likely to exist in the Paradox Member.

Hydrochemical data for the Paradox Member support interpretations based on potentiometric-head data. According to Thackston and others (1981, p. 215), water in the Paradox Member is either a sodium chloride or calcium chloride type with substantial concentrations of magnesium and sulfate. The dissolved-solids concentration in this water ranges from 6,700 to 440,000 mg/L (Rush and others, 1982, p. 45; Weir and others, 1983a, p. 48 and 1983b, p. 40; Whitfield and others, 1983, p. 49; INTERA Environmental Consultants, Inc., 1984, p. 61), with no discernible geographical trends (Thackston and others, 1981, p. 215). The salinity of the water and the apparent absence of geographical trends in the dissolved-solids concentration indicate that water in the Paradox Member largely is isolated from any active regional flow system (Thackston and others, 1981, p. 215).

Vertical flow through the Paradox Member, however, is suggested by differences in potentiometric head between overlying and underlying aquifers (for example, see Rush and others, 1982, p. 13; Weir and others, 1983a, p. 29) and by the similar chemistry of water in the Paradox Member and underlying Mississippian carbonate rocks in the Four Corners aquifer system. According to Hanshaw and Hill (1969, p. 285), this similarity indicates that the Four Corners aquifer system in the Paradox Basin is recharged primarily by water percolating down through the Paradox Member. Nearly all investigators in the Paradox Basin agree that vertical flow is restricted to discrete discontinuities, such as faults, joints, stratigraphic pinch-outs, the margins of igneous intrusions, or the edges of salt diapirs (for example, see Rush and others, 1982, p. 12; Whitfield and others, 1983, p. 21). Springs issuing from the Paradox Member, including Tripp and Trimble Hot Springs (Barrett and Pearl, 1977, p. 241), Stinking Spring (Ritzma and Doelling, 1969, p. 106–107), and Onion Creek Spring (Rouse, 1967, p. 19), represent either water rising along faults in areas having upward head gradients or meteoric water descending through fractures in outcrops near which the springs issue. Typically, these springs have discharges of less than 100 gal/min.

Similar to the Paradox Member in the Paradox Basin, the Eagle Valley Evaporite in the Eagle Basin, White River Plateau, and Elk Mountains contains water traveling mostly along local flow paths. All known springs and wells that discharge water from the Eagle Valley Evaporite are located in stream valleys; with one exception, the concentration of dissolved solids in this water ranges from 250 to 2,600 mg/L (Rouse, 1967, p. 14; Brogden and Giles, 1976a; Giles and Brogden, 1976). Considering the topographic setting and chemical quality, most of the water appears to originate as precipitation on nearby outcrops; some may be seepage from channel alluvium.

Some of the water in the Eagle Valley Evaporite is water in transit between the Canyonlands aquifer and Four Corners aquifer system. This water has been in circulation longer and consequently is more saline than water typically discharging from the formation to springs and shallow wells. As an example, Big Spring at Dotsero, Colo., has a dissolved-solids concentration of 10,700 mg/L (Iorns and others, 1964, p. 678–679). The water issuing from Big Spring probably originated on the White River Plateau, descended through the Leadville Limestone and other aquifers, then rose to the surface at the base of the uplift along flanking faults. As noted previously, saline springs discharging from the Leadville Limestone at Glenwood Springs, Colo., are attributable to downward percolation of meteoric water through the Eagle

Valley Evaporite in uplifted areas to the south. As in the Paradox Member of the Hermosa Formation, vertical movement through the Eagle Valley Evaporite would be negligible without the existence of discrete discontinuities, such as faults and joints.

Extrapolating from the above information, it would appear that subregional flow in conjunction with the Canyonlands aquifer occurs in the upper part of the Four Corners confining unit, particularly in areas beyond the depositional extent of evaporite facies in the Eagle Valley Evaporite and Paradox Member of the Hermosa Formation. Within areas containing thick deposits of shale and evaporites, flow is dominated by water in transit between the Canyonlands aquifer and the Four Corners aquifer system along faults, joints, and other vertical discontinuities and by local, topographically controlled flow paths. In areas characterized by extensive faulting, such as the Verdure and Lisbon Valley grabens in Utah or the Cattle Creek graben in Colorado, considerable exchange of water between aquifers above and below the Four Corners confining unit can occur (URS Corporation, 1982, 1983; INTERA Environmental Consultants, Inc., 1984, p. 139–143).

CIRCULATION IN THE CANYONLANDS AQUIFER

Flow in the Canyonlands aquifer is greatly influenced by topography and facies changes within hydrogeologic units that make up the aquifer, more so than in the Four Corners aquifer system. Numerous small to moderately sized springs issuing from the Canyonlands aquifer in uplifted areas within a few miles of where water enters the aquifer (fig. 97) indicate that, in the Canyonlands aquifer, local flow paths are at least as important as subregional flow paths in the movement of ground water.

As in the Four Corners aquifer system, the Canyonlands aquifer is demarcated by the Continental Divide from the San Juan Mountains to the Sierra Madre and potentiometric divides elsewhere. In places, the potentiometric divides do not coincide with the political boundaries of the UCRB (pl. 11). These potentiometric divides are the Wind River–Northern Great Plains divide on the north, the Upper Colorado River Basin–Basin and Range divide on the west, the San Juan–Monument divide on the southeast, and according to Cooley and others (1969, pl. 5), a divide possibly in the vicinity of the Mogollon Slope (location shown on pl. 1) on the south. Within the UCRB, potentiometric divides affecting the movement of water in the Canyonlands aquifer include the Uinta-Park and Tavaputs divides.

North of the Uinta-Park divide, water in the Canyonlands aquifer moves from peripheral highlands and the internal Rock Springs Uplift to the edges of structural basins, where it discharges to springs, is pumped from wells, or rises into Mesozoic and Tertiary rocks. Moderate to large springs, such as springs on La Barge Creek (SB29–117–01ad) and Hams Fork (SB26–18–13bad), and a few water wells deplete much of the water recharged in the Overthrust Belt within a few miles of where the water enters the Canyonlands aquifer. Pinching out of the geologic units that comprise the Canyonlands aquifer a few miles south of Kendall, Wyo., and disruption of the aquifer by the Wind River fault have produced warm and cold springs at Kendall (fig. 98) with a combined discharge of about 8 ft^3/s. These springs probably represent all of the water entering the Canyonlands aquifer from the Gros Ventre Range and Wind River Mountains along the northern boundary of the UCRB. Wells and springs, such as Olson Spring, in the Rawlins area also deplete water from the Canyonlands aquifer, and oil wells in the Great Divide Basin and Rock Springs Uplift (pl. 11 and table 20) withdraw large quantities of water from the aquifer. A cone of depression more than 1,000 ft deep apparently exists around the Lost Soldier–Wertz–Mahoney oil fields (pl. 11). Most of the water not pumped from wells or discharged from springs probably rises into Mesozoic or Tertiary rocks in the Green River, Great Divide, and Washakie Basins because of large upward head differences in these areas (pl. 12). Ultimately, some ground water flows out of the UCRB into the Hanna Basin through a gap between the Rawlins Uplift and Sierra Madre.

Between the Uinta-Park and Tavaputs divides, water in the Canyonlands aquifer moves from peripheral and internal highlands toward the Yampa and Green Rivers. Ground water recharged on the east side of the area flows toward the Piceance and Eagle Basins and then moves northwesterly to the Yampa River valley (Roger L. Hoeger, consulting hydrologist, written commun., 1969). Numerous small to moderately sized springs, such as Conundrum Hot Spring (Barrett and Pearl, 1977), intercept water recharged in the mountainous regions. In addition, numerous water wells completed mostly in the Maroon Formation in and near Glenwood Springs, Colo., discharge water from the Canyonlands aquifer at a combined rate of about 900 gal/min (Wright Water Engineers, 1979). Ground water entering the Canyonlands aquifer on the west side of the area flows into the interior of the Uinta Basin and then moves northeast and east along the valleys of the Green and Duchesne Rivers toward the Yampa River (Hood and Fields, 1978, p. 34). Several large to very large springs, such as Warm Springs (UB01–08–30ddb), Ratliff Spring (SLD02–22–31adc), and Ashley Creek Springs (SLD03–20–01d), drain the Canyonlands aquifer at the base of the Uinta Mountains (Maxwell and others, 1971, p. 15; Hood and others, 1976, p. 34–35). Also, water wells completed mostly in the Weber Sandstone in and near the Uinta Mountains discharge water from the aquifer at a combined rate of about 2,700 gal/min. By far, the largest pumpage from the Canyonlands aquifer between the Uinta-Park and Tavaputs divides comes from the Rangely and Ashley Valley oil fields. The combined water production from these two oil fields in December 1986 was nearly 36 ft^3/s (table 20). Water not pumped from wells or discharged to springs along local flow paths rises into Mesozoic and Tertiary rocks in the

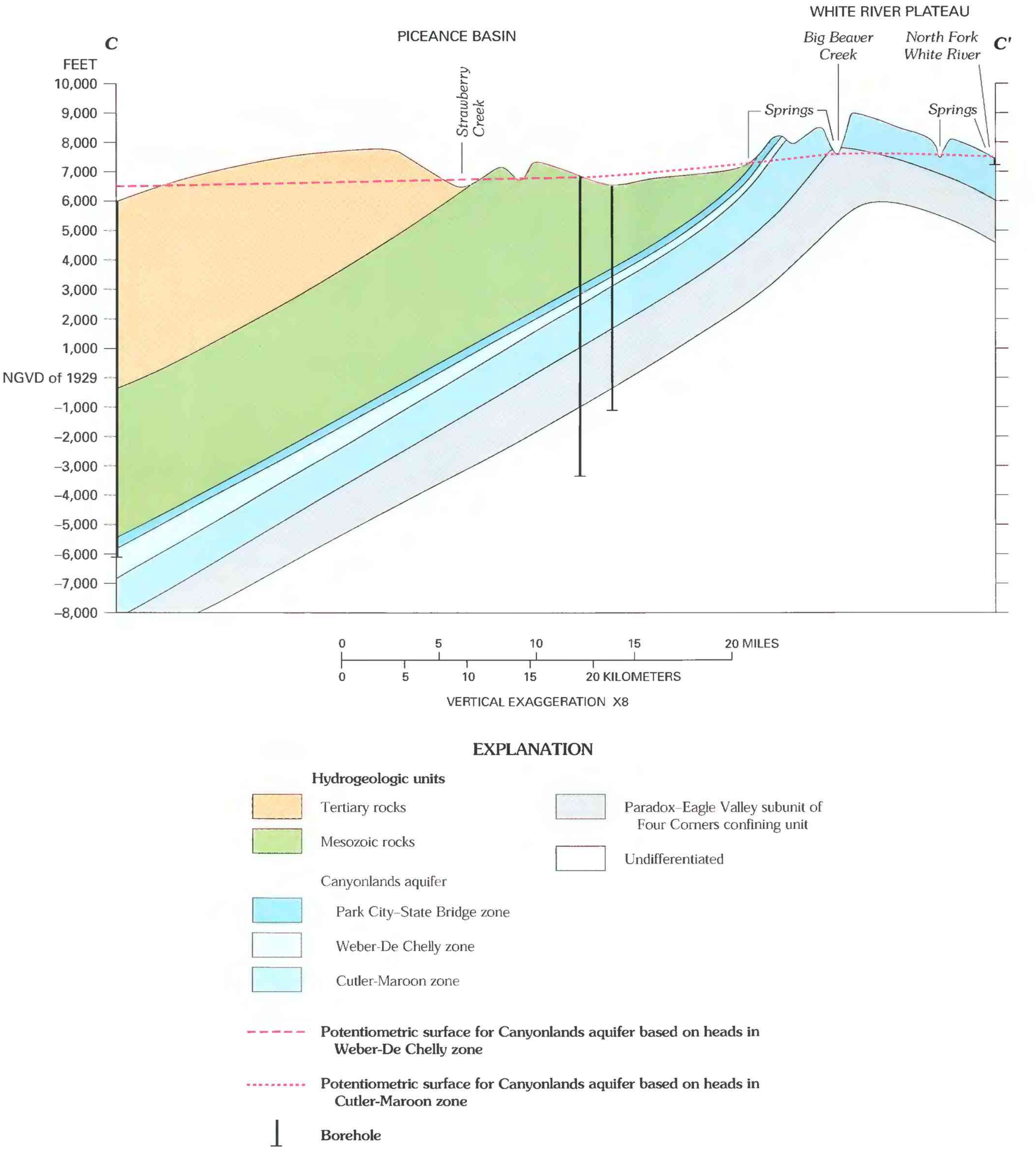


FIGURE 97.—Geologic section from the Piceance Basin to the White River Plateau indicating the potentiometric surface for the Canyonlands aquifer and potential spring locations (location of section shown on pl. 11).

FIGURE 98.—Kendall Warm Springs (SB39–110–36) issuing from the Phosphoria Formation and flowing over terrace deposits into the Green River north of Pinedale, Wyoming.

vicinity of the Green and Yampa Rivers or flows toward the Jones Hole area at the eastern end of the Uinta Mountains (pl. 11). The combined flow of all springs in Jones Hole is 37 ft^3/s (Hood and others, 1976, p. 34–35).

South of the Tavaputs divide, water movement in the Canyonlands aquifer is influenced more strongly by topography than in any other part of the UCRB. West of the Green and Colorado Rivers, water in the Canyonlands aquifer moves from the San Rafael Swell, Circle Cliffs Uplift, High Plateaus, and Kaibab Plateau toward the valleys of the Price, San Rafael, Dirty Devil, Escalante, and Paria Rivers (Hood and Patterson, 1984, p. 31–33). Between the Green and Colorado Rivers and the San Juan–Monument divide, water in the Canyonlands aquifer flows from the Uncompahgre Plateau, San Juan Mountains, Monument Upwarp, and Defiance Plateau toward the valleys of the Colorado, Green, Dolores, and San Juan Rivers (Cooley and others, 1969, pl. 5; Richter, 1980, fig. 10; Weir and others, 1983b, pl. 2; INTERA Environmental Consultants, Inc., 1984).

Some recharge to the Canyonlands aquifer south of the Tavaputs divide occurs around the edges of laccolithic mountains, whereas grabens in the area can be conduits for groundwater movement to or from the aquifer. In the Henry Mountains, for example, meteoric water infiltrates the Canyonlands aquifer along the fractured margins of igneous intrusions (Hood and Danielson, 1981, p. 32–36). Ultimately, however, this water ascends under a moderately upward hydraulic gradient into Mesozoic rocks in the Henry Mountains Basin (pl. 12). The Abajo Mountains, La Sal Mountains, and the intervening Sage Plain are considered to be a major recharge area for the Canyonlands aquifer in the Paradox Basin (Thackston and others, 1981; Weir and others, 1983b; Ackerman and Rush, 1984). This recharge may be augmented by upward leakage of water from the Four Corners aquifer system between the Paradox Valley and Sinbad Valley grabens because salt deposits normally present between the Canyonlands aquifer and Four Corners aquifer system have been plastically deformed and are thin or absent between Paradox Valley and Sinbad Valley (Weir and others, 1983b, p. 26). Gains in streamflow by the Dolores River near Gateway, Colo., and by the Colorado River in the Salt Valley and Spanish Valley grabens might be, in part, also related to upward leakage. Conversely, leakage from the Canyonlands aquifer to the Four Corners aquifer system probably is occurring in the Big Gypsum Creek, Lisbon Valley, and Verdure grabens (Richter, 1980, p. 38; INTERA Environmental Consultants, Inc., 1984, p. 165).

South of the Tavaputs divide, numerous small to moderately sized springs drain the Canyonlands aquifer wherever canyons are entrenched into plateau surfaces. At the western edge of the Uncompahgre Plateau, for example, Placerville Warm Spring issues from the Cutler Formation where the San Miguel River is incised into the plateau (fig. 99). Abetted by the large permeability of rocks that form most of the surface area in the Canyonlands region, numerous seeps and springs issue from the Cedar Mesa and White Rim Sandstones not far from where water infiltrates these formations (Sumsion and Bolke, 1972; Huntoon, 1977; Hand, 1979; Richter, 1980; Rush and others, 1982). One such spring (SLD31–20–30add) occurs at the base of Angel Arch in the Needles district of Canyonlands National Park. The largest spring in the Monument Upwarp, Big Spring (SLD32–18–29dbd), discharges into Gypsum Canyon at a rate of 125 gal/min (Huntoon, 1979). At the edges of the Defiance Plateau, numerous small to moderately sized springs issue from the De Chelly Sandstone (Davis and others, 1963).

FIGURE 99.—Placerville Warm Spring (NMB44–11–34ddd) issuing from travertine deposits at the base of an outcrop of the Cutler Formation at Placerville, Colorado. The spring location appears to be controlled by intersecting faults. On August 15, 1988, the date the spring was photographed, its discharge was 3.5 gallons per minute.

South of the Tavaputs divide and west of the San Juan–Monument divide, very little water is withdrawn from the Canyonlands aquifer by wells. The only oil or gas fields in the area with significant water production from the Canyonlands aquifer, Upper Valley and Boundary Butte, discharged water at a combined rate of 643 gal/min in December 1986 (table 20). A few domestic and stock wells are completed in the Canyonlands aquifer in the San Miguel River valley (Ackerman and Brooks, 1985), near Moab, Utah (Sumsion, 1971), in Canyonlands National Park and Natural Bridges National Monument (Sumsion and Bolke, 1972; Huntoon, 1977), and in the Defiance Plateau area (Davis and others, 1963; McGavock and others, 1966).

All water in the Canyonlands aquifer south of the Tavaputs divide, west of the San Juan–Monument divide, and north of the Mogollon Slope that is not discharged locally to wells or springs either rises into Mesozoic rocks in structurally low areas, such as the Castle Valley Sag, Paradox Basin, or Chinle Wash area (pl. 12), or flows out of the UCRB into canyons of the Colorado River (Marble Canyon) and Little Colorado River south of Lees Ferry, Ariz. A few small springs issue from the Coconino Sandstone and Supai Group in these canyons (Metzger, 1961, p. 128), but according to Cooley (1976, p. 8–9), the primary mechanism of discharge from the Canyonlands aquifer south of Lees Ferry is the downward movement of water along faults and fractures to the Four Corners aquifer system. Therefore, some of the water in springs issuing from the Four Corners aquifer system in Marble Canyon and the canyon of the Little Colorado River, such as Blue Spring, represents subregional discharge from the Canyonlands aquifer.

RATES OF GROUND-WATER MOVEMENT

If geochemical effects on ground-water movement, such as osmotic effects, density gradients, chemical reactions between the water and host rock, adsorption, and diffusion are ignored, rates of lateral ground-water movement can be estimated for two regional water-bearing zones, the Redwall-Leadville zone of the Madison aquifer and the Weber-De Chelly zone of the Canyonlands aquifer, based on the following equation:

$$V = \frac{TI}{\Phi b} \tag{25}$$

where

V = average linear velocity, in feet per day;
T = transmissivity, in feet squared per day;
I = hydraulic gradient, in feet per foot;
Φ = porosity, dimensionless; and
b = thickness of aquifer, in feet.

Rates of ground-water movement through the Redwall-Leadville and Weber-De Chelly zones represent maximum values for all Paleozoic rocks in the UCRB because no other hydrogeologic units are more permeable.

In the Redwall-Leadville zone of the Madison aquifer, lateral ground-water velocities are estimated to range from 0.000001 to 0.001 ft/d in structural basins and from 0.001 to 600 ft/d on the flanks and in the centers of uplifted areas (pl. 13). Rates of movement in excess of 10 ft/d are estimated to occur only in the southern part of the White River Plateau, where the rocks are cavernous and extensively faulted and fractured. At estimated rates of movement in this hydrogeologic unit, it would take ground water between 14,500 and 14,500,000 years to travel 1 mi laterally in structural basins but only between 9 days and 14,500 years to travel 1 mi laterally in and on the flanks of uplifted areas.

In the Weber-De Chelly zone of the Canyonlands aquifer, lateral ground-water velocities are estimated to range from 0.000005 to 0.001 ft/d in structural basins and from 0.001 to 2 ft/d on the flanks and in the centers of uplifted areas (pl. 13). It would take water moving through this hydrogeologic unit between 14,500 and 2,900,000 years to travel 1 mi laterally in structural basins but only between 7.2 and 14,500 years to travel 1 mi laterally in and on the flanks of uplifted areas.

In general, estimated average linear velocities indicate that water moving through the Paleozoic rocks along subregional flow paths from upland recharge areas to structural basins is slowed by decreasing transmissivity and hydraulic gradients in the direction of movement. In the center of structural basins, lateral ground-water movement essentially stagnates, and the prevailing direction of movement is upward into Mesozoic and Tertiary rocks.

DISCHARGE FROM THE PALEOZOIC ROCKS

Water in the Paleozoic rocks discharges to streams, springs, and wells, is consumed by evapotranspiration where the rocks are at or near land surface, seeps up into Mesozoic and Tertiary rocks in structural basins, or flows out of the UCRB into adjacent hydrologic basins. Discharge areas for the Paleozoic rocks are shown in figure 92. The greatest potential for discharge probably occurs in areas where head differences between the Canyonlands aquifer and the Four Corners aquifer system are more negative than –500 ft (pl. 12). These areas include valleys within the Overthrust Belt; the Green River, Great Divide, Washakie, Sand Wash, Eagle, Piceance, Blanding, Paradox, and San Juan Basins; topographically low parts of the Uinta Basin; the southern Henry Mountains Basin; upper Chinle Valley; Castle Valley; the San Rafael Desert; the Wasatch Plateau, lower reaches of the Yampa, Duchesne, Dolores, Green, Dirty Devil, and San Juan Rivers; Marble Canyon; and the canyon of the Little Colorado River.

SEEPAGE TO STREAMS AND SPRINGS

In areas where the Paleozoic rocks are at or near land surface, discharges to streams, springs, and seeps account for the largest ground-water outflow (see, for example, Hood and Fields, 1978). Discharges to springs and seeps commonly occur along local flow paths by one of four mechanisms: (1) Intersection of the land surface and potentiometric surface; (2) prevention of downward ground-water movement by negligibly permeable layers; (3) downdip changes in permeability within aquifers; or (4) fault severing of aquifers.

Most commonly, local discharge to springs and streams develops where deep canyons are incised below the potentiometric surface of an aquifer (fig. 97). Springs and seeps issue from fractures and caverns etched into cliffs bordering the canyons. Representative springs include Vasey's Paradise Spring in Marble Canyon (Cooley, 1976, p. 7); numerous springs in canyons draining the White River Plateau (Iorns and others, 1964; Teller and Welder, 1983); and saline springs issuing along the Colorado River in Cataract Canyon (Thackston and others, 1981, p. 208). Discharges from these springs can fluctuate substantially in response to seasonal and annual variations in precipitation. The discharge of Vasey's Paradise Spring, for example, ranges from 45 to 4,500 gal/min.

Ground water discharges locally from plateaus, buttes, and mesas in the Colorado Plateaus province where downward percolating ground water encounters a formation or layer with small to negligible permeability and is forced to flow laterally toward bounding escarpments. Such contact springs are numerous throughout the Monument Upwarp, on the flanks of the Defiance Plateau, and in the Grand Canyon (fig. 100).

On the homoclinal edges of uplifted areas, including the Rawlins Uplift, San Rafael Swell, Circle Cliffs Uplift, and Monument Upwarp, springs can develop in response to changes in permeability within an aquifer (Huntoon, 1983, p. 18–29). Typically, as aquifers become less fractured and more cemented away from the axes of uplifts, hydraulic conductivity decreases by several orders of magnitude. Water flowing downdip is rejected and either flows upward into overlying formations or discharges at springs (fig. 101).

Along the thrust-faulted margins of mountain ranges, including the southwestern flank of the Wind River Mountains and Gros Ventre Range, the north and south flanks of the Uinta Mountains, and the east flanks of the Salt River and Wyoming Ranges, aquifers typically are severed by faults, producing separate circulation systems in the hanging wall and foot wall (fig. 101). Large springs typically develop in the hanging wall where aquifers are thrust against formations with negligible permeability and water is forced to rise along the fault to the surface. Representative of this type of spring, Hogsback Spring on La Barge Creek near La Barge, Wyo., issues from the Lodgepole and Mission Canyon Limestones in the hanging wall of the Darby-Hogsback Thrust Fault (Lines and Glass, 1975). Measured discharges from this spring range from 4,000 to 5,500 gal/min (Lines and Glass, 1975; U.S. Geological Survey, unpub. data).

In most areas where the Paleozoic rocks are at or near land surface, spring discharges and seepage to channel alluvium along subregional and local flow paths enhance the discharge of through-flowing streams. Measured gains in streamflow in reaches where Paleozoic rocks compose all or

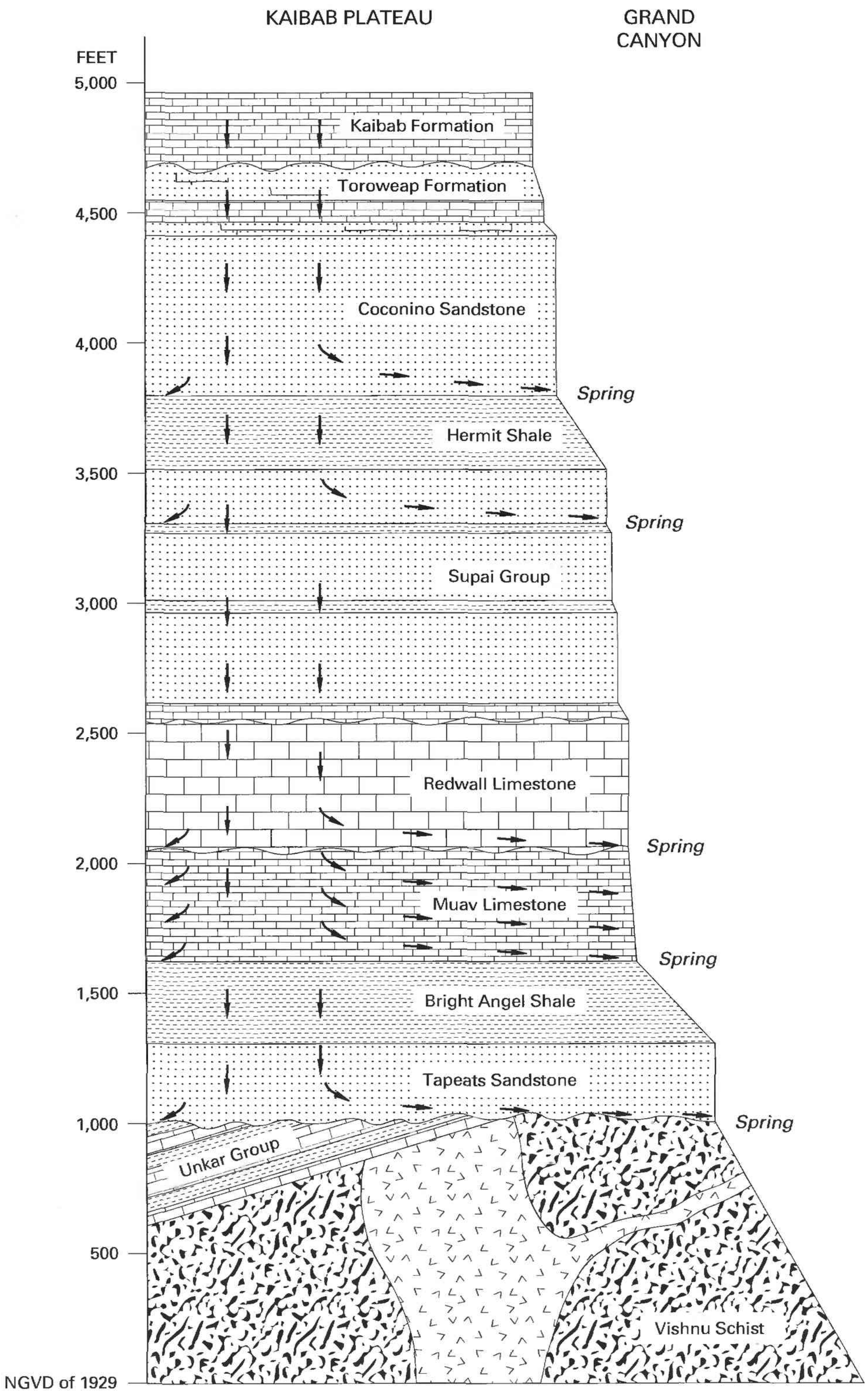


FIGURE 100.—Schematic representation of ground-water movement with respect to stratigraphy in the Kaibab Plateau in Arizona. (Modified from Metzger, 1961, pl. 14.)

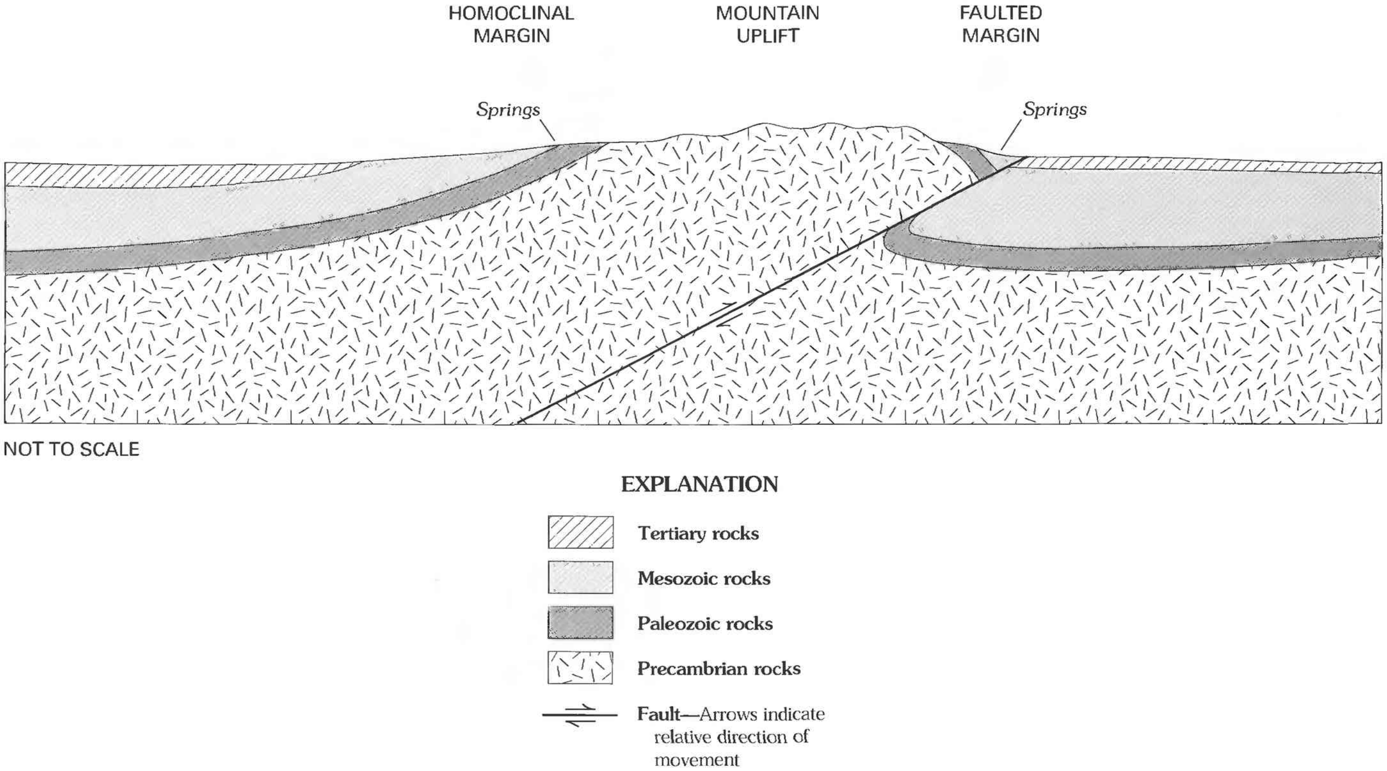


FIGURE 101.—Schematic section showing potential spring locations on the flanks of an uplifted area. (Modified from Huntoon, 1983, p. 14.)

most of the bedrock range from 0.61 to 7.0 ft^3/s/mi; most gains range from 1 to 3 ft^3/s/mi (table 21). Streams with measured gains in discharge from Paleozoic rocks include the Colorado, Green, White, Blue, Eagle, Crystal, Fryingpan, Roaring Fork, San Rafael, Dolores (excluding Big Gypsum valley), and Uncompahgre Rivers (Rush and others, 1982, p. 34–42; Weir and others, 1983b, p. 34; Warner and others, 1985). The Duchesne River, where it leaves the western Uinta Mountains, is believed to gain streamflow from Paleozoic, Mesozoic, and Tertiary rocks (Hood, 1977, p. 14). The San Juan River, where it cuts through Paleozoic rocks in the Monument Upwarp, and the Yampa River, where it is incised into Paleozoic rocks in the eastern Uinta Mountains, also are believed to be gaining streams (Cooley and others, 1969, p. 44; Sumsion, 1976, p. 41). Small streams with perennial flow attributable to ground-water discharge from Paleozoic rocks include Maroon, Gypsum, Sweetwater, Deep, and Rifle Creeks (Teller and Welder, 1983; Warner and others, 1985). Total ground-water outflows to springs and streams from Paleozoic rocks within the Upper Colorado River Basin are known to be at least 1,113 ft^3/s or about 810,000 acre-ft/yr (table 22). Undoubtedly, additional springs and seeps exist but have not been recorded.

WITHDRAWAL FROM WELLS

Water production from wells completed in the Paleozoic rocks is insignificant in most areas of the UCRB. In uplifted areas, where these rocks are at or near land surface, scattered wells provide water for domestic or stock use, usually at rates of less than 50 gal/min. However, a few of these wells discharge water at rates of 100 to 3,000 gal/min. Most of this production occurs in the McCoy, Aspen, Crested Butte, Telluride, Ouray, and Glenwood Springs areas of Colorado (Bryant, 1972; Hampton, 1974; Wright Water Engineers, 1979; Giles, 1980; Ackerman and Brooks, 1985; City of Ouray, written commun, 1988); in and near the Uinta Mountains (Hood, 1976; Hood and others, 1976; Sumsion, 1976); in the Overthrust Belt (Lines and Glass, 1975) and Rawlins Uplift (Berry, 1960); in and near the Defiance Plateau (Davis and others, 1963; McGavock and others, 1966; Levings and Farrar, 1977a); and in the Paradox Basin and Canyonlands areas of southeastern Utah (Feltis, 1966; Ritzma and Doelling, 1969; Sumsion, 1971; Sumsion and Bolke, 1972; Hand, 1979; Richter, 1980; Weir and others, 1983a; Hood and Patterson, 1984). The total potential discharge of all water wells completed in the Paleozoic rocks as of 1988, that could be determined from published and unpublished data, is about 50 ft^3/s, or about 36,000 acre-ft/yr (table 22). Because some of these wells were not in operation when visited between 1984 and 1988 and because others may be operating at less than their reported capacities, the current (1988) production of water from these wells is assumed to be less than 36,000 acre-ft/yr.

The largest withdrawals of water from Paleozoic rocks occur in conjunction with oil and gas production. As shown in tables 20 and 22, oil and gas wells completed mostly in Paleozoic reservoirs in the Rangely, Lost Soldier–Wertz–Mahoney, greater Aneth, Ashley Valley, Upper Valley, Brady, greater Lisbon, and 23 smaller fields or consolidated fields produced water at a combined rate of about 63 ft^3/s, or about 46,000 acre-ft/yr, in December 1986 (data from Petroleum Information Corporation, written commun., 1987). Actual water production from the Paleozoic rocks alone is unknown but has to be less than 46,000 acre-ft/yr because some production from these fields is from Mesozoic rocks.

TABLE 21.—*Measured ground-water outflows to streams from Paleozoic rocks in the Upper Colorado River Basin, excluding the San Juan Basin*

Stream	Reach	Channel geology	Base flow			Inferred discharge from Paleozoic rocks	Source of data
			Gain (cubic feet per second)	Length of reach (miles)	Unit gain (cubic feet per second per mile)		
		Carbonate and clastic rocks					
Colorado River	Glenwood Canyon, Colo.	Sawatch Quartzite to Belden Formation; some Precambrian rocks	30	17	1.8	All	URS Corporation (1983)
Eagle River	Headwaters to Avon, Colo.	Sawatch Quartzite to Minturn Formation some Precambrian rocks and till	83	38	2.2	More than half	Warner and others (1985)
Sweetwater Creek	Headwaters to Colorado River	Leadville Limestone to Eagle Valley Evaporite	27	16	1.7	All	Warner and others (1985)
Deep Creek	Headwaters to Colorado River	Sawatch Quartzite to Belden Formation; minor Precambrian rocks	11	14	.79	All	Warner and others (1985)
Rifle Creek	Huffman Gulch to Rifle Falls, Colo.	Leadville Limestone to Maroon Formation	28	4	7.0	All	Teller and Welder (1983)
		Mostly clastic rocks					
Colorado River	Cisco, Utah, to mouth of Green River	Hermosa, Rico, and Cutler Formations and Mesozoic rocks	300	97	3.1	More than half	Rush and others (1982)
Colorado River	Mouth of Green River to Hite, Utah	Hermosa Formation and Cutler Group	95	54	1.8	All	Rush and others (1982)
Colorado River	Kremmling to Dotsero, Colo.	Eagle Valley Evaporite to State Bridge Formation; some Mesozoic and Precambrian rocks	34	56	.61	Most	Warner and others (1985)
Crystal River	Headwaters to Carbondale, Colo.	Mostly Eagle Valley Evaporite to Maroon Formation; some Leadville Limestone, Mesozoic and igneous rocks and till	84	42	2.0	More than half	Warner and others (1985)
North Fork White River	Trappers Lake to Buford, Colo.	Mostly Morgan and Maroon Formations and Weber Sandstone; some stream and glacial alluvium	62	12	5.2	Most	Boyle and others (1984)
Dolores River	Headwaters to Rico, Colo.	Mostly Hermosa, Rico, and Cutler Formations; some Mesozoic and Precambrian rocks	25	20	1.3	More than half	Warner and others (1985)
Dolores River	Gateway, Colo., to Cisco, Utah	Cutler Formation and Mesozoic rocks	46	19	2.4	About half	Warner and others (1985)
Maroon Creek	Above Aspen Highlands Ski area, Colo.	Mostly Maroon Formation; some Mesozoic and igneous rocks	35	17	2.1	Most	Warner and others (1985)
		Mostly shale and evaporites					
Eagle River	Avon to Gypsum, Colo.	Mostly Eagle Valley Evaporite; some Maroon Formation and Mesozoic rocks	34	22	1.6	More than half	Warner and others (1985)
Gypsum Creek	Headwaters to Gypsum, Colo.	Mostly Eagle Valley Evaporite; some Maroon Formation and till	34	30	1.1	More than half	Warner and others (1985)

TABLE 22.—*Summary of known discharges from Paleozoic rocks to streams, springs, and wells in the Upper Colorado River Basin, excluding the San Juan Basin, as of 1988*

[≤, less than or equal to; ≥, greater than or equal to]

Area	Four Corners aquifer system and Four Corners confining unit		Canyonlands aquifer	
	Discharge point	Discharge (cubic feet per second)	Discharge point	Discharge (cubic feet per second)
North of Uinta Mountains and Axial Basin Arch	Springs in the Overthrust Belt	27.2	Springs in the Overthrust Belt	5.6
	Springs in the Rawlins Uplift	.2	Springs in the Rawlins Uplift	.4
	Sheep Creek Spring (Uinta Mountains)	6.5	Kendall Springs (Wind River Mountains)	7.8
	Water wells in the Rawlins Uplift	1	Water wells in the Overthrust Belt	1
	Miscellaneous gas-field pumpage[1]	.4	Lost Soldier–Wertz–Mahoney oil-field pumpage[1]	19.2
			Miscellaneous oil- and gas-field pumpage	.9
	Subtotal (rounded)	35		35
Between Uinta Mountains and Uncompahgre Plateau	Miscellaneous springs in the Uinta Mountains[2]	41.5	Miscellaneous springs in the Uinta Mountains[2]	13.9
	Split Mountain Warm Spring (Uinta Mountains)	6.0	Jones Hole Springs (Uinta Mountains)	37.0
	Springs in the Elk Mountains	.9	Springs in the Elk Mountains	2.4
	Bowles Hatchery Spring (Sawatch Range)	2.5	Seepage to Crystal River	[3]≥19
	Springs and seeps in Glenwood Canyon	30.2	Seepage to North Fork White River	[3]≥47
	Springs on the White River Plateau	5.8	Seepage to Colorado River	[3]≥14
	Seepage to Rifle Creek	27.7	Seepage to Maroon Creek	[3]≤35
	Seepage to Sweetwater Creek	[3]≤27	Seepage to Eagle River	[3]≤8
	Seepage to Deep Creek	[3]11	Seepage to Yampa River	[3]≤109
	Seepage to Eagle River	[3]≤46	Water wells in and near the Uinta Mountains	6
	Water wells in the McCoy area	[4]25	Water wells in the Glenwood Springs area	2
	Water wells in the Glenwood Springs area	[5]4.5	Rangely oil-field pumpage	33.5
			Ashley Valley oil-field pumpage	2.4
	Subtotal (rounded)	228		329
South of Uncompahgre Plateau	Springs in the San Juan Mountains	2.2	Springs in the San Juan Mountains and Paradox Basin	.6
	Water wells at Ouray, Colo.	1.5	Springs on the Monument Upwarp	1.5
	Aneth oil-field pumpage	5.0	Springs on the Defiance Plateau	.4
	Lisbon oil- and gas-field pumpage	.4	Seepage to Colorado River	[3]≥304
	Miscellaneous oil- and gas-field pumpage	.1	Seepage to Green River	[3]≤54
			Seepage to San Juan River	[3]≤171
			Seepage to Dolores River	[3]≥47
			Seepage to Dirty Devil and San Rafael Rivers	[3]±1
			Water wells in San Miguel River Valley	.1
			Water wells in southeastern Utah	4
			Water wells in the Defiance Plateau area	5
			Miscellaneous oil- and gas-field pumpage	1.5
	Subtotal (rounded)	9		590
	Total (rounded)	272		954

[1]Water production from oil and gas fields is for December 1986. Production rates may not be representative of average 1986 production because of monthly fluctuation. Production figures were obtained from Petroleum Information Corporation (written commun., 1987) for fields indicated on State maps to be producing mostly from Paleozoic reservoirs. Some fields produce from both the Four Corners aquifer system and the Canyonlands aquifer (both in Paleozoic rocks), and some produce from Paleozoic and Mesozoic rocks. Listings in this table reflect the dominant producing horizon.

[2]Springs originating from streamflow losses to sinks immediately upstream are not included.

[3]Estimate based on ground-water seepage rates and approximate length of stream bordered by formations in either the Four Corners aquifer system or Canyonlands aquifer.

[4]Reported discharges from three wells not known to be in use in 1987.

[5]Average discharge of Redstone 21–9 geothermal well in 1984 aquifer test and Wright well. Both wells were not in use in 1987.

EVAPOTRANSPIRATION

Evapotranspiration is estimated to consume a large amount of the ground water in stream valleys and negligible amounts elsewhere. Because of a generally shallow water table along streams, water there is easily evaporated from soil, alluvium, and the uppermost layers of bedrock and is taken up readily by plants. Evapotranspiration around springs and in stream valleys where Paleozoic rocks are at or near the surface has been estimated in various studies to consume ground water at rates of between 0.5 and 2 ft^3/s per square mile of phreatophyte cover (table 23). In the northern Uinta Basin, more ground water is lost to evapotranspiration than is discharged to springs and streams (Hood and Fields, 1978).

LEAKAGE TO MESOZOIC AND TERTIARY ROCKS

Upward leakage of water from Paleozoic to Mesozoic and Tertiary rocks occurs around the edges and in the interior of structural basins (see Hampton, 1974). The largest proportion of upward movement occurs along discrete vertical discontinuities, such as faults, joints, stratigraphic pinch-outs, or the margins of igneous intrusions or salt diapirs (Rush and others, 1982, p. 12; Whitfield and others, 1983, p. 21; INTERA Environmental Consultants, 1984, p. 139–143). Ultimately, ground water rising in structural basins discharges to through-flowing streams. The Green, Colorado, Yampa, Dolores, San Juan, and Dirty Devil Rivers are the principal loci for upward ground-water movement in structural basins (pl. 12). Because of uncertainties in the vertical hydraulic-conductivity values and vertical hydraulic gradients in the Paleozoic rocks of the UCRB, the rate of upward leakage from the Paleozoic rocks cannot be estimated. However, because the structural basins in which this leakage is believed to occur generally are arid, it is estimated that leakage of water from Paleozoic rocks to Mesozoic and Tertiary rocks is the principal mechanism of recharge to the Mesozoic and Tertiary rocks in these areas.

INTERBASIN FLOW

Peripheral divides and hydraulic gradients prevent ground-water outflow from the study area in all but three places. Between the Rawlins Uplift and Sierra Madre, ground water flows out of the Great Divide Basin and into the Hanna Basin, part of the Northern Great Plains regional aquifer system (see Downey, 1984). In the southeastern corner of the study area, ground water flows under the Four Corners Platform into the San Juan Basin, which is part of the UCRB but was studied separately. While outflows to both the Hanna and San Juan Basins are believed to represent a small proportion of the annual ground-water discharge from the Paleozoic rocks of the UCRB, substantial quantities of ground water flow beneath the Kaibab Plateau and Marble Platform toward the confluence of the Colorado and Little Colorado Rivers (INTERA Environmental Consultants, Inc., 1984, p. 59), which is in the Lower Colorado River Basin.

According to Cooley (1976, p. 8), the combined flow of all springs and seeps in the Little Colorado River where it flows into the Colorado River ranged from 217 to 232 ft^3/s and averaged 223 ft^3/s between 1950 and 1967. Some of this water represents local recharge from the Kaibab and Coconino Plateaus and Marble Platform or recharge from the San Francisco Plateau and Mogollon Rim (in the Lower Colorado River Basin), but a considerable amount of the water originates in the Defiance Plateau and Monument Upwarp areas of the UCRB (Cooley, 1976, p. 13).

TABLE 23.—*Ground-water consumption by evapotranspiration and seepage in areas where Paleozoic rocks are at or near land surface in the Upper Colorado River Basin*

[<, less than]

Location	Drainage area (square miles)	Area covered by phreatophytes (square miles)	Ground-water outflow: Evapotranspiration (acre-feet per year)	Ground-water outflow: Evapotranspiration (cubic feet per second per square mile)	Ground-water outflow: Springs and seeps (acre-feet per year)	Source of data
Paradox Basin						
Blanding-Durango area	4,600	21.9	27,000	1.7	<200	Whitfield and others (1983, p. 47)
Dolores River Basin	3,000	57.4	45,000	1.1	110,000	Weir and others (1983b, p. 37)
Moab-Monticello area	1,900	23.4	32,000	1.9	210,000	Weir and others (1983a, p. 44)
Green River-Moab area	3,000	39.1	24,000	.86	81,000	Rush and others (1982, p. 44)
Henry Mountains Basin						
Lower Dirty Devil River Basin	4,300	93.8	30,000	.44	75,000	Hood and Danielson (1981, p. 44)
Uinta Mountains and Uinta Basin						
South flank of Uinta Mountains and northern Uinta Basin	5,200	347	160,000	.64	130,000	Hood and Fields (1978, p. 18–19)

Springs issuing from the Paleozoic rocks in Marble Canyon have a combined discharge of 11 ft^3/s, most of which comes from Vasey's Paradise Spring and two other springs (McCulley, 1985, p. 13). According to McCulley (1985), the outflow from springs in Marble Canyon is a mixture of water recharged west of the Colorado River on the Kaibab Plateau and water moving southwesterly from the Kaiparowits Basin.

The combined flow of all springs issuing from Paleozoic rocks in the canyon of the Little Colorado River and in Marble Canyon averages 234 ft^3/s (170,000 acre-ft/yr). The preceding discussion indicates that a considerable amount of this springflow appears to be outflow from the UCRB to the Lower Colorado River Basin.

GROUND-WATER BUDGET FOR THE UPPER COLORADO RIVER BASIN

The ground-water budget for the UCRB cannot be determined accurately because of considerable uncertainty in determining several of the water-budget components. What is known is summarized in table 24. Resolution of uncertainties awaits additional data collection. At present, only a generalized model of the southern UCRB has been done by Weiss (1990). Detailed modeling of the entire UCRB is needed to verify or revise the information presented in this report.

TABLE 24.—*Estimated ground-water budget for the Paleozoic rocks of the Upper Colorado River Basin, excluding the San Juan Basin*

[<, less than; >, greater than]

Water-budget component	Annual volume (acre-feet)
Inflows	
Infiltration of precipitation	<6,600,000
Interbasin ground-water flow	1,000
Outflows	
Springs and seepage	>810,000
Well withdrawals	
Water wells	<36,000
Oil and gas wells	<46,000
Evapotranspiration	Unknown
Leakage to Mesozoic and Tertiary rocks	Unknown
Interbasin flow	
To Hanna Basin	Unknown
To San Juan Basin	Unknown
To Lower Colorado River Basin	<170,000

SUMMARY AND CONCLUSIONS

The Upper Colorado River Basin (UCRB), excluding parts of the San Juan Basin, is an area of about 100,000 mi^2. The UCRB is included in four physiographic provinces—the Middle Rocky Mountains, Wyoming Basin, Southern Rocky Mountains, and Colorado Plateaus. Numerous uplifts and structural basins occur within each province, segmenting the UCRB into a variety of landforms, including mountains, plateaus, cuestas, hogbacks, mesas, plains, badlands, intermontane basins, river valleys, and canyons. The general surface of the UCRB lies at altitudes between 5,000 and 8,000 ft, but mountains and plateaus rise to altitudes as high as 14,500 ft, and streams are entrenched to altitudes as low as 3,100 ft.

In general, precipitation is distributed orographically. Average annual precipitation ranges from less than 6 inches in lowlands to more than 60 inches in some mountains but it fluctuates by more than 15 inches from year to year in some areas. Average precipitation exceeds 10 in/yr at altitudes above 5,900 ft. At lower altitudes, most of the precipitation is consumed by evapotranspiration. At higher altitudes, most of the precipitation between April and September is consumed by evapotranspiration. Within the part of the UCRB covered by this study, precipitation averages about 15 in/yr.

The Colorado and Green Rivers are the principal streams in the UCRB. At its confluence with the Green River, the Colorado River has an average discharge of about 8,000 ft^3/s, about 1,400 ft^3/s more than the average discharge of the Green River, even though the Green River at the confluence is longer and has a larger drainage area. Tributaries of the Colorado and Green Rivers with average discharges in excess of 500 ft^3/s include the San Juan, Gunnison, Yampa, Roaring Fork, Dolores, White, New Fork, Duchesne, and Eagle Rivers. Eighty-two reservoirs in the basin have a combined storage capacity of 38 million acre-ft, 71 percent of which is in Lake Powell. Between 1963 and 1982, an average annual reduction of 5,680 ft^3/s in the discharge of the Colorado River where the river leaves the UCRB is attributable to construction of Glen Canyon Dam and Lake Powell. Leakage of water from Lake Powell into adjacent Paleozoic and Mesozoic rocks caused ground-water levels near the lake to rise by as much as 500 ft between 1963 and 1983.

Paleozoic rocks in the UCRB were classified in this report into 11 hydrogeologic units on the basis of the predominant lithologic and hydrologic characteristics of sedimentary sequences that persist throughout most of the basin. However, each of these sequences is heterogeneous because of various proportions of included rock types and various degrees of fracturing, solution channeling, cementation, and compaction. In general, all hydrogeologic units are most permeable and, therefore, have the largest sustained yields to wells and springs in uplifted areas and are least permeable in the interiors of structural basins. However, within comparable structural settings, there is little relation between permeability and depth of burial. The permeability of sandstone appears to be closely related to porosity and, hence, grain size and the degree of cementation. In contrast, the permeability of carbonate rocks generally is not related to porosity but depends more on the presence of fractures and solution openings. Ranges overlap, but relative to confining units, aquifers characteristically have larger unit-averaged porosity, permeability, and hydraulic conductivity, larger composite transmissivity, and larger sustained yields to wells and springs.

The lower six hydrogeologic units composed of Paleozoic rocks constitute the Four Corners aquifer system. At the bottom of this aquifer system, the Flathead aquifer contains as much as 800 ft of friable to firmly cemented sandstone and quartzite with minor carbonate rocks, shale, and conglomerate. Gradationally overlying the Flathead aquifer, the Gros Ventre confining unit consists of as much as 1,100 ft of sandy shale with subordinate sandstone and carbonate rocks. Above the Gros Ventre confining unit, the Bighorn aquifer consists of as much as 3,000 ft of limestone and dolomite with subordinate shale and minor sandstone. Above the Bighorn aquifer, the Elbert-Parting confining unit consists of as much as 700 ft of variably interbedded carbonate rocks, sandstone, quartzite, shale, and anhydrite.

The uppermost aquifer in the Four Corners aquifer system is the Madison aquifer, which includes the Redwall-Leadville zone and the overlying Darwin-Humbug zone. The Redwall-Leadville zone, which is as much as 2,500 ft thick, consists of limestone and dolomite, with minor sandstone and shale, and sparse to abundant chert layers, fragments, and nodules. Dolomite in the Redwall-Leadville zone increases in proportion to limestone basinward. The Darwin-Humbug zone, which is as much as 800 ft thick and limited in distribution, consists of variable proportions of limestone, dolomite, sandstone, shale, solution breccia, and gypsum.

The Redwall-Leadville zone of the Madison aquifer is the most transmissive hydrogeologic unit in the Four Corners aquifer system, and, consequently, hydrologic properties of this zone represent maximum values for the aquifers and confining units that make up the Four Corners aquifer system. Unit-averaged porosity in the Redwall-Leadville zone ranges from less than 1 to 11 percent; local-scale permeability (permeability determined by aquifer tests or calculated from laboratory-determined values) ranges from 0.02 to 1,800 md; unit-averaged hydraulic conductivity ranges from 0.00005 to 200 ft/d; composite transmissivity ranges from less than 0.01 to 47,000 ft^2/d; storativity where the aquifer is at least 100 ft thick is estimated to range from 0.005 to 0.0002. Artesian (natural) yields from wells and springs typically range from less than 1 to 10,000 gal/min.

Between the Four Corners aquifer system and the uppermost hydrogeologic units composed of Paleozoic rocks is the Four Corners confining unit, which in ascending order, includes the Belden-Molas and Paradox–Eagle Valley subunits. The Belden-Molas subunit, which is as much as 4,300 ft thick, consists of shale with subordinate carbonate rocks and sandstone, and minor gypsum. The Paradox–Eagle Valley subunit generally is less than 9,700 ft thick, but small diapiric intrusions in the Eagle and Paradox Basins are estimated to be as much as 15,000 ft thick. The Paradox–Eagle Valley subunit consists of various proportions of carbonate rocks, shale, sandstone, gypsum-anhydrite, and halite. Unit-averaged porosity in the Four Corners confining unit ranges from less than 1 to 6 percent; local-scale permeability ranges from 0.0012 to 550 md; unit-averaged hydraulic conductivity ranges from 0.00001 to 0.03 ft/d; composite transmissivity ranges from 0.001 to 50 ft^2/d. Artesian yields from wells and springs typically range from less than 1 to 150 gal/min.

The upper three hydrogeologic units composed of Paleozoic rocks form the Canyonlands aquifer, which in ascending order, includes the Cutler-Maroon, Weber-De Chelly, and Park City–State Bridge zones. The Cutler-Maroon zone, which is as much as 16,500 ft thick, consists of highly variable proportions of arkosic sandstone, quartz sandstone, conglomerate, shale, carbonate rocks, and gypsum-anhydrite. The Weber-De Chelly zone, which may be as much as 4,000 ft thick, consists of friable to firmly cemented quartz sandstone with minor shale and carbonate rocks. The Park City–State Bridge zone, which is as much as 800 ft thick, consists of highly variable proportions of shale, carbonate rocks, sandstone, gypsum-anhydrite, phosphorite, and chert.

Within the Canyonlands aquifer, the Weber-De Chelly zone transmits most of the water, but where this zone pinches out south and east of the Colorado River and north of the San Juan River, the Cutler-Maroon zone is the most transmissive part of the Canyonlands aquifer. Within the Canyonlands aquifer, unit-averaged porosity ranges from 1 to 28 percent; local-scale permeability ranges from 0.00078 to 380 md; unit-averaged hydraulic conductivity ranges from 0.00001 to 20 ft/d; composite transmissivity ranges from 0.00005 to 10,000 ft^2/d; and artesian yields from wells and springs range from less than 1 to 2,900 gal/min. Storativity in the Weber-De Chelly zone where it is more than 100 ft thick is estimated to range from 0.00003 to 0.001.

Recharge to the Paleozoic rocks occurs in peripheral and interior highlands primarily from precipitation on outcrops, infiltration of meteoric water through overlying strata, or leakage of runoff from stream channels. Recharge is possible in areas receiving as little as 10 inches of precipitation annually. The total direct and indirect recharge from precipitation is estimated to be less than 6,600,000 acre-ft/yr. On the western and northern edges of the UCRB, additional recharge is provided by ground-water inflows from structurally continuous areas separated from the UCRB by Quaternary surface-water divides. This recharge is estimated to be about 1,000 acre-ft/yr. In general, recharge areas are characterized by predominantly downward ground-water movement, losing streams, unsaturated rock, water-table conditions, and relatively fresh ground water. The largest potential for recharge exists where head differences between the Canyonlands and Madison aquifers are the most positive. The principal recharge areas, on this basis, are the Wind River, Uinta, Elk, Abajo, and Elkhead Mountains; La Sal Mountains; the Sierra Madre, Park, and Sawatch Ranges; the Uncompahgre, Kaibito, Kaibab, and White River Plateaus; the High Plateaus region; topographically high areas in the Uinta and Piceance Basins; the Verdure, Lisbon Valley, and Big Gypsum Creek grabens in the Paradox Basin; and an arcuate area extending from the San Juan Mountains through the Sage Plain to the Monument Upwarp.

Because of highly variable topography within and peripheral to the UCRB, water in the Paleozoic rocks of the UCRB is forced to flow toward local and subregional outlets, rather than toward a single, regional discharge area. Although some water flows vertically through the Four Corners confining unit, the generally small permeability of this confining unit inhibits interchange of water between the underlying Four Corners aquifer system and the overlying Canyonlands aquifer. Within the Four Corners aquifer system and Canyonlands aquifer, water generally moves from structural and erosional highlands toward structural and fluvial basins. Ground-water movement in the Paleozoic rocks ultimately is directed mainly toward four areas—the eastern Great Divide Basin (between the Rawlins Uplift and Sierra Madre), the confluence of the Yampa and Green Rivers, the San Juan Basin, and the confluence of the Colorado and Little Colorado Rivers. The fourth area is in the Lower Colorado River Basin, just south of the UCRB.

Average linear velocities of water in the Redwall-Leadville zone of the Madison aquifer and the Weber-De Chelly zone of the Canyonlands aquifer can be estimated from the distributions of composite transmissivity, hydraulic gradient, thickness, and unit-averaged porosity in these hydrogeologic units. Rates of lateral ground-water movement in the Redwall-Leadville zone of the Madison aquifer are estimated to range from 0.000001 to 600 ft/d. At these rates, it would take between 14,500,000 years and about 9 days for water to travel 1 mi laterally. Rates of lateral ground-water movement in the Weber-De Chelly zone of the Canyonlands aquifer are estimated to range from 0.000005 to 2 ft/d. At these rates, it would take between 2,900,000 and 7.2 years for water to travel 1 mi laterally. Average linear velocities decrease from uplifted areas to structural basins as a consequence of decreasing hydraulic gradients and transmissivity basinward.

Water in the Paleozoic rocks discharges to streams, springs, and wells, is consumed by evapotranspiration where the rocks are at or near land surface, seeps upward into Mesozoic and Tertiary rocks in structural basins, or flows out of the UCRB into adjacent hydrologic basins. In general, discharge areas are characterized by upward ground-water movement, gaining streams, springs, flowing artesian wells, and brackish to briny ground water. The largest potential for discharge exists where head differences between the Canyonlands and Madison aquifers are the most negative. The principal discharge areas, on this basis, are the greater Green River Basin, the San Juan Basin, the southeastern Piceance and Uinta Basins, the eastern Paradox Basin, the southwestern Henry Mountains Basin, the upper Chinle Valley, the San Rafael Desert, Castle Valley, and the lower reaches of the Colorado, Green, Yampa, Dolores, Dirty Devil, and Little Colorado Rivers.

In areas where Paleozoic rocks are at or near land surface, discharges to springs and streams are the largest ground-water outflow. Discharges to springs and streams commonly occur along local flow paths by one of four mechanisms: (1) Intersection of the topographic and potentiometric surfaces, (2) prevention of downward percolation by negligibly permeable layers, (3) downdip changes in permeability within aquifers, or (4) fault severing of aquifers. Springflow and seepage along subregional and local flow paths augment streamflow at rates commonly between 1 and 3 ft^3/s per mile of channel bordered by Paleozoic rocks. Streams with measured or inferred gains in discharge from Paleozoic rocks include the Colorado, Green, San Juan, Yampa, White, Dolores, Crystal, Fryingpan, Roaring Fork, Eagle, Blue, Dirty Devil, San Rafael, and Uncompahgre Rivers and Maroon, Sweetwater, Deep, Gypsum, and Rifle Creeks. Total known outflows to springs and streams within the UCRB are estimated to equal or exceed about 810,000 acre-ft/yr.

Water production from Paleozoic rocks is negligible in most areas. In uplifted areas, where the rocks are at or near land surface, scattered wells provide water for domestic and stock use at rates of generally less than 50 gal/min, but a few wells discharge water at rates of 100 to 3,000 gal/min. If all water wells known to be completed in the Paleozoic rocks were in operation in 1988, their total production would be about 36,000 acre-ft/yr. Considerable water production also occurs from oil and gas fields in which the principal reservoir is the Paleozoic rocks. The largest of these fields are the Rangely, Lost Soldier–Wertz–Mahoney, Aneth, Upper Valley, Brady, and Ashley Valley fields. In December 1986, the 30 oil and gas fields in the UCRB that produced mostly from reservoirs in the Paleozoic rocks yielded water at a combined rate of about 46,000 acre-ft/yr.

Evapotranspiration is estimated to consume a large amount of ground water in stream valleys and negligible amounts elsewhere. Evapotranspiration around springs and in stream valleys where Paleozoic rocks are at or near land surface consumes ground water at estimated rates of 0.5 to 2 ft^3/s per square mile of phreatophyte cover.

Leakage of ground water from the Paleozoic rocks to Mesozoic and Tertiary rocks is believed to occur around the edges and in the interior of structural basins, principally in the vicinity of vertical discontinuities such as faults, stratigraphic pinch-outs, or the edges of igneous intrusions or salt diapirs. This leakage, which could not be quantified because of insufficient data, may be the principal mechanism of recharge to Mesozoic and Tertiary rocks in the interior of deep, generally arid structural basins within the UCRB.

Peripheral ground-water divides and hydraulic gradients prevent ground-water outflow from the study area in all but three places. Between the Rawlins Uplift and Sierra Madre, ground water flows out of the Great Divide Basin into the Hanna Basin. Ground water also flows between the Monument Upwarp and Defiance Plateau beneath the Four Corners Platform and into the San Juan Basin (a part of the UCRB that was excluded from this study). A third flow path from the UCRB directs water under the Kaibab Plateau and Marble Platform to the confluence of the Colorado and Little Colorado Rivers in the Lower Colorado River Basin. Outflows to the Hanna and San Juan Basins are unknown but are believed to be a small part of the annual ground-water budget for the Paleozoic rocks. However, springs issuing from Paleozoic rocks near the confluence of the Colorado and Little Colorado Rivers have a combined average discharge of 170,000 acre-ft/yr. Although some of this springflow represents water recharged locally and water moving along subregional flow paths from the Lower Colorado River Basin, a considerable amount of this springflow is outflow from the UCRB.

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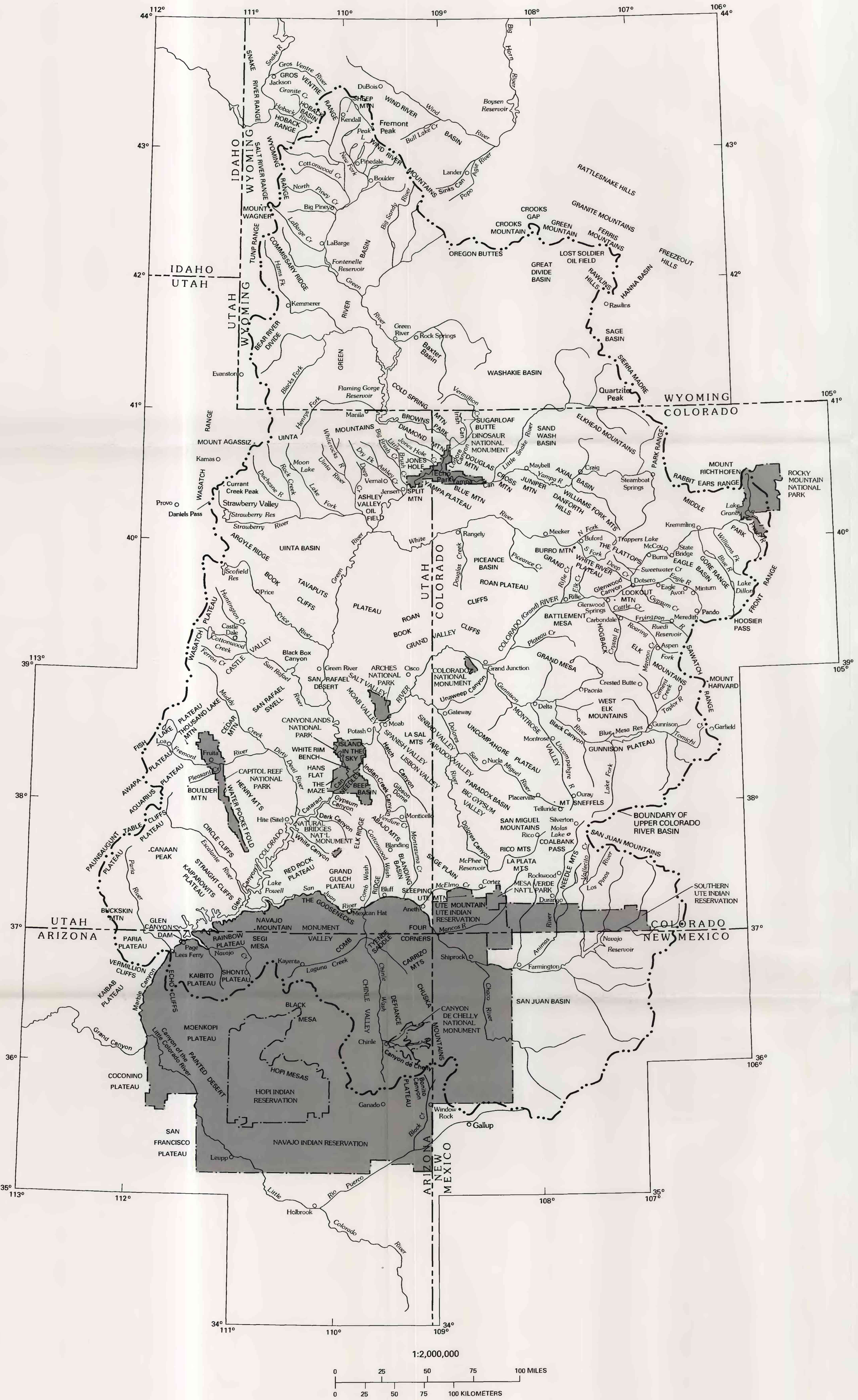
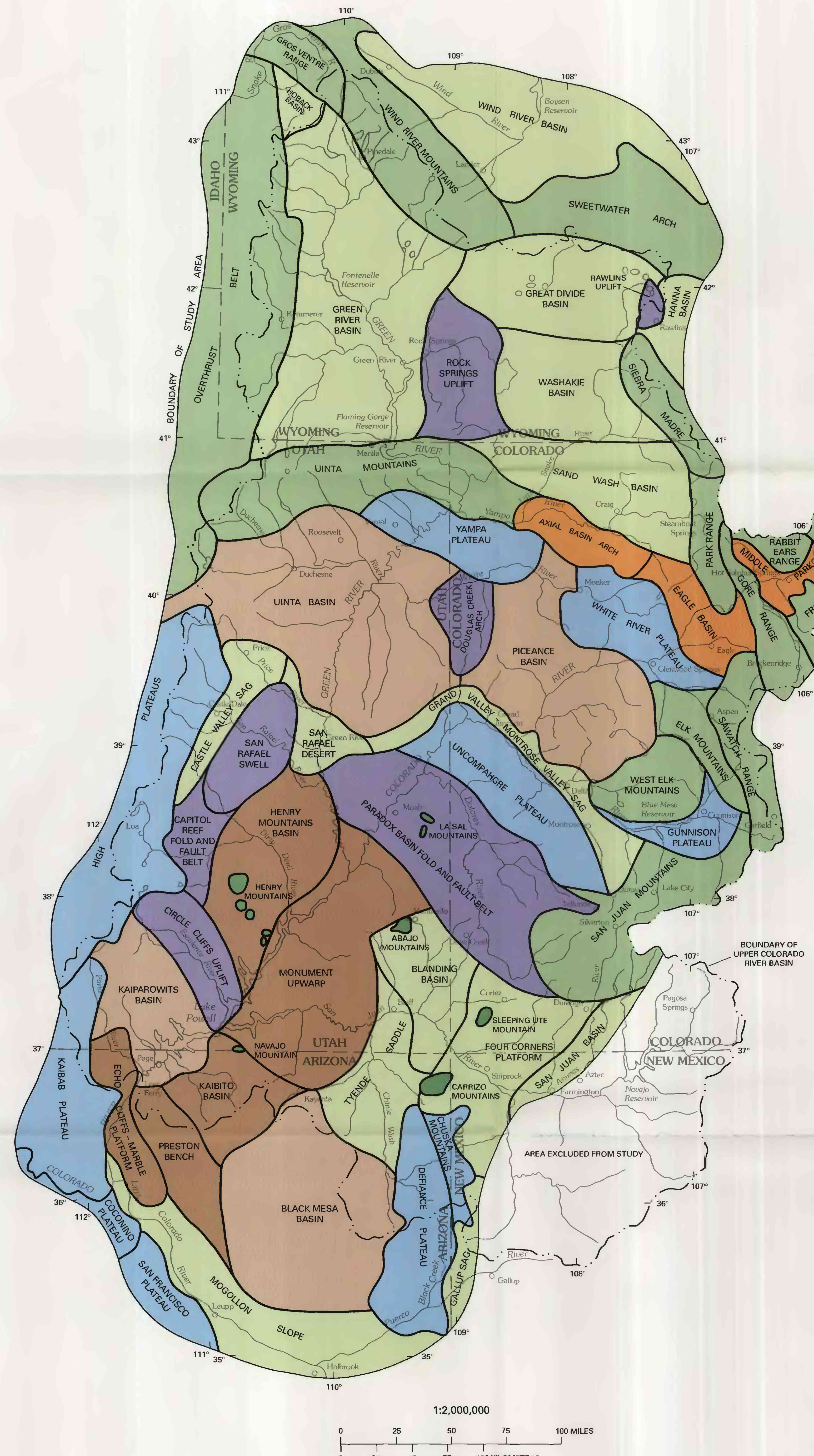
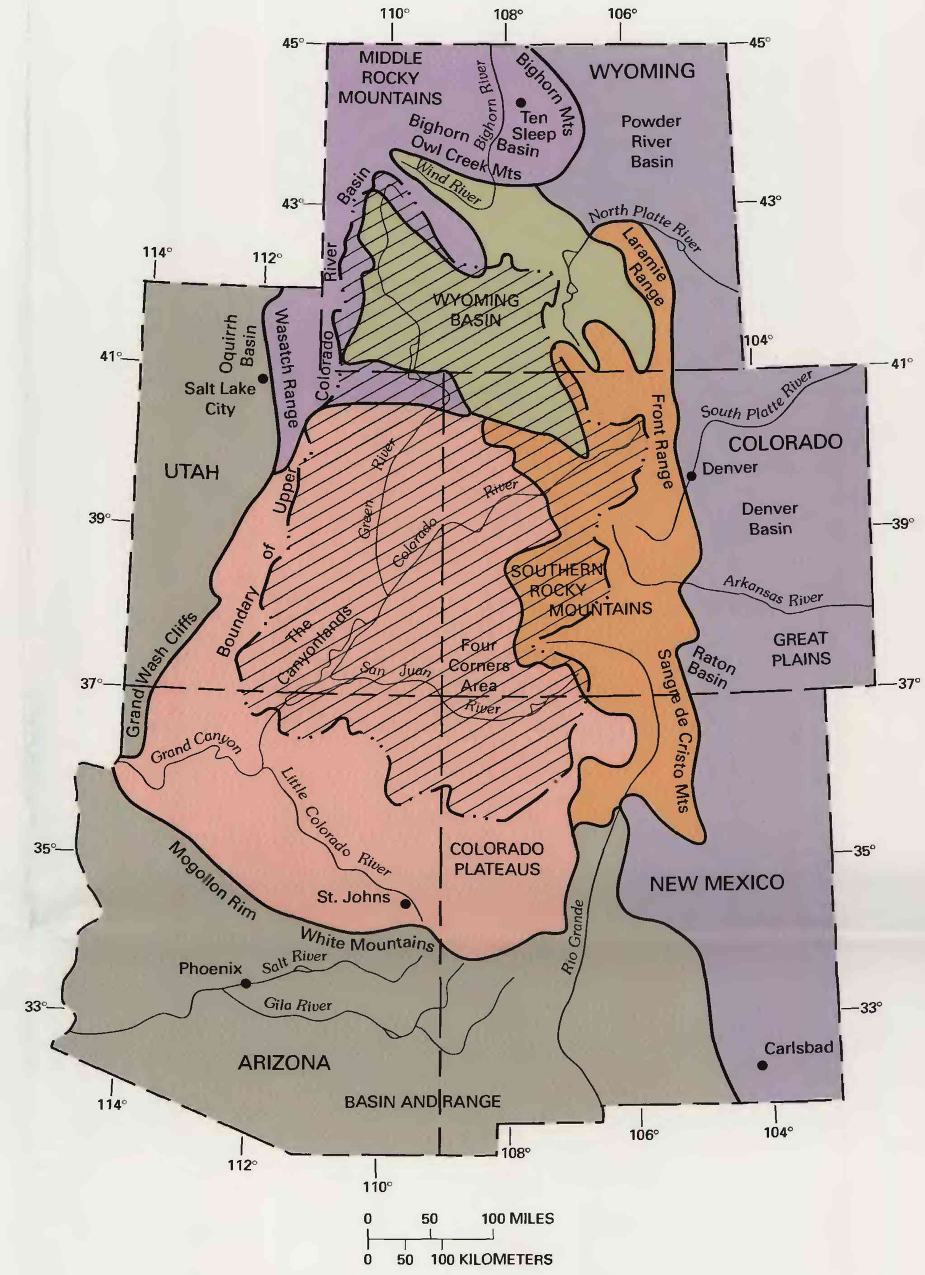
PROFESSIONAL PAPER 1411–B
Geographic features—PLATE 1
Geldon, A.L., 2002, Hydrologic properties and ground-water flow systems of the Paleozoic rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, excluding the San Juan Basin



CITIES, RIVERS, PHYSICAL FEATURES, NATIONAL PARKS, AND INDIAN RESERVATIONS



TOPOGRAPHY



INDEX MAP SHOWING LOCATION OF UPPER COLORADO RIVER BASIN WITH RESPECT TO PHYSIOGRAPHIC PROVINCES

EXPLANATION

Topography

Mountain ranges—Peak altitudes range from 10,000 to 14,000 feet above NGVD of 1929

Isolated mountains—Peak altitudes range from 9,000 to 12,000 feet above NGVD of 1929

High plateaus—Surface altitudes range from 8,000 to 12,000 feet above NGVD of 1929

Broad, dissected plateaus—Surface altitudes range from 6,500 to 11,000 feet above NGVD of 1929 on plateau rims and from 5,000 to 6,500 feet in interior of plateaus. Plateau escarpments are ruggedly dissected and 1,000 to 4,000 feet high

Stair-stepped plateaus—Incised by canyons and surmounted by buttes and mesas. Surface altitudes generally between 4,000 and 8,000 feet above NGVD of 1929

Irregular topography—Lowlands flanked by ridges, cuestas, and mesas that are incised by deep canyons. Altitudes range from 4,500 to 9,500 feet above NGVD of 1929

Rolling plains, hills, and low mountains—Surface altitudes generally 6,000 to 8,500 feet above NGVD of 1929. Hills and mountains rise 1,000 to 3,000 feet above plains

Plains, badlands, and mesas with isolated hills and mountains—Surface altitudes range from 6,000 to 7,500 feet above NGVD of 1929 north of the Uinta and Piceance Basins; from 4,000 to 7,000 feet above NGVD of 1929 south of the Uinta and Piceance Basins

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

GEOGRAPHIC FEATURES OF THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

By
Arthur L. Geldon
2002

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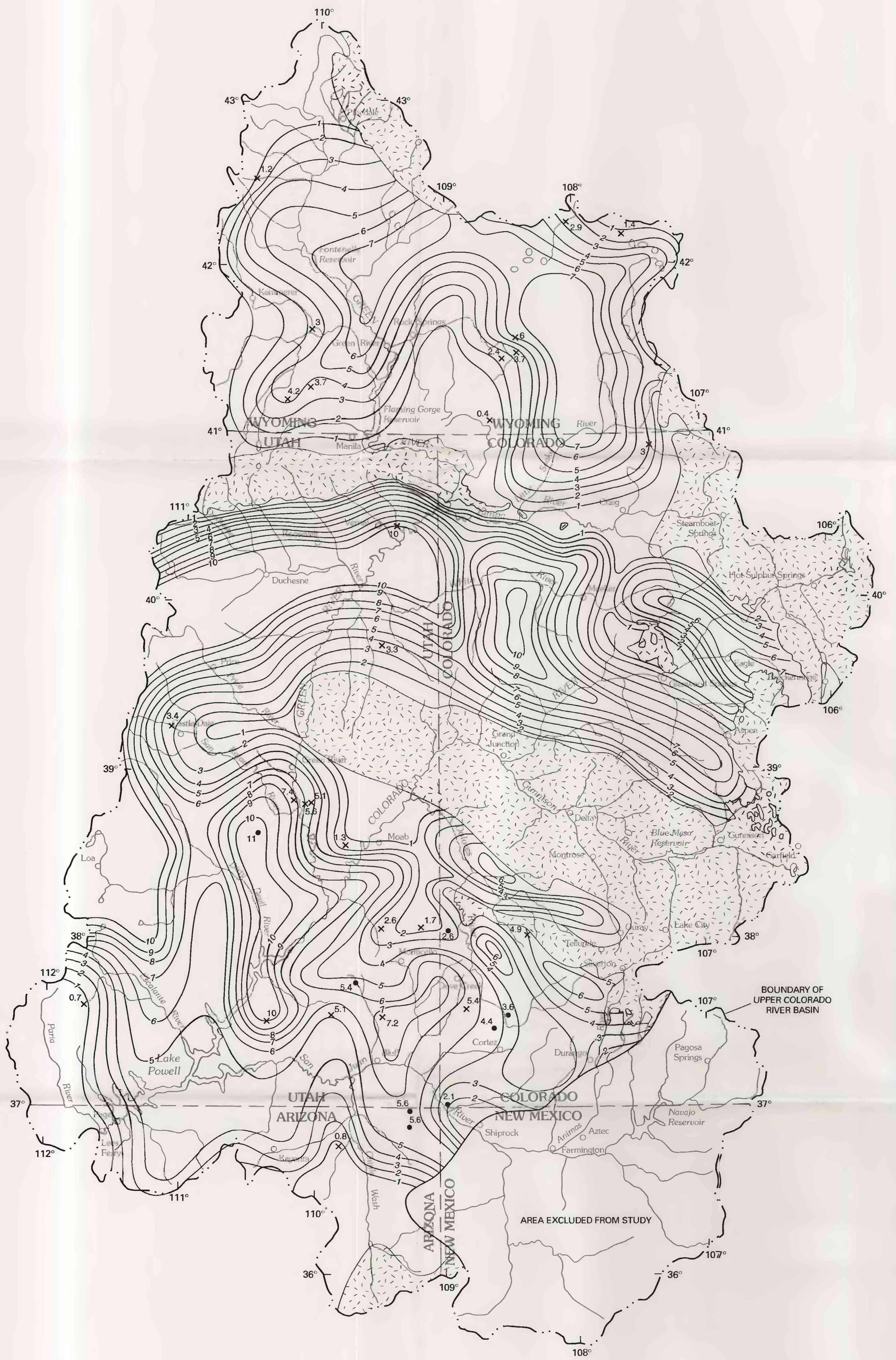
PROFESSIONAL PAPER 1411–B
Porosity and hydraulic conductivity of the Redwall-Leadville zone—PLATE 2
Geldon, A.L., 2002, Hydrologic properties and ground-water flow systems of the Paleozoic rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, excluding the San Juan Basin



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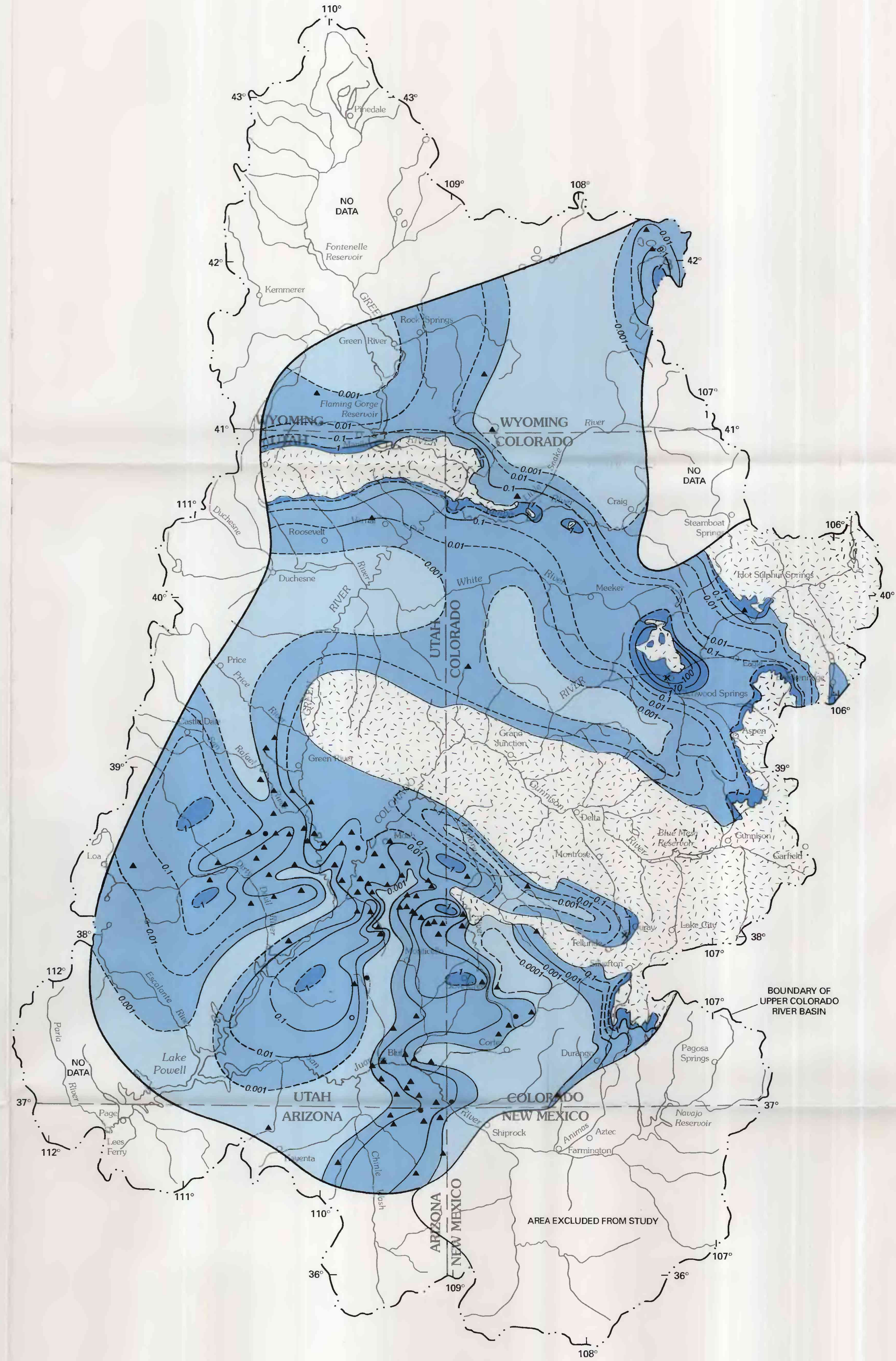
0 25 50 75 100 MILES

0 25 50 75 100 KILOMETERS

UNIT-AVERAGED POROSITY



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UNIT-AVERAGED HYDRAULIC CONDUCTIVITY

EXPLANATION

Area where the Redwall-Leadville zone is missing because of erosion or nondeposition

Unit-Averaged Porosity

— 5 — **Line of equal estimated unit-averaged porosity**—Location is approximate. Interval is 1 percent

• 4.4 **Site at which unit-averaged porosity was estimated as the mean of laboratory-determined values in a borehole interval at least 100 feet thick and representative of the lithology of the Redwall-Leadville zone at the site**—Number shown is mean porosity in the interval, in percent

× 5.4 **Site at which unit-averaged porosity was estimated as the median of geophysically determined values in a borehole interval at least 100 feet thick**—Number shown is median porosity in the interval, in percent

Unit-Averaged Hydraulic Conductivity

Relative unit-averaged hydraulic conductivity

- Large
- Moderate
- Small

— 0.01 — — **Line of equal estimated unit-averaged hydraulic conductivity**—Dashed where approximately located. Interval, in feet per day, is variable

Limit of data

Data sites

▲ Site at which estimate was based on permeability determined by a drill-stem test

• Site at which estimate was based on the mean of laboratory-determined permeability values in a borehole interval

○ Site at which estimate was based on the average of hydraulic-conductivity values determined by pressure-injection tests in a borehole interval

× Site at which estimate was based on transmissivity determined by a pumping or flowing-well test

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

UNIT-AVERAGED POROSITY AND HYDRAULIC CONDUCTIVITY OF THE REDWALL-LEADVILLE ZONE OF THE MADISON AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

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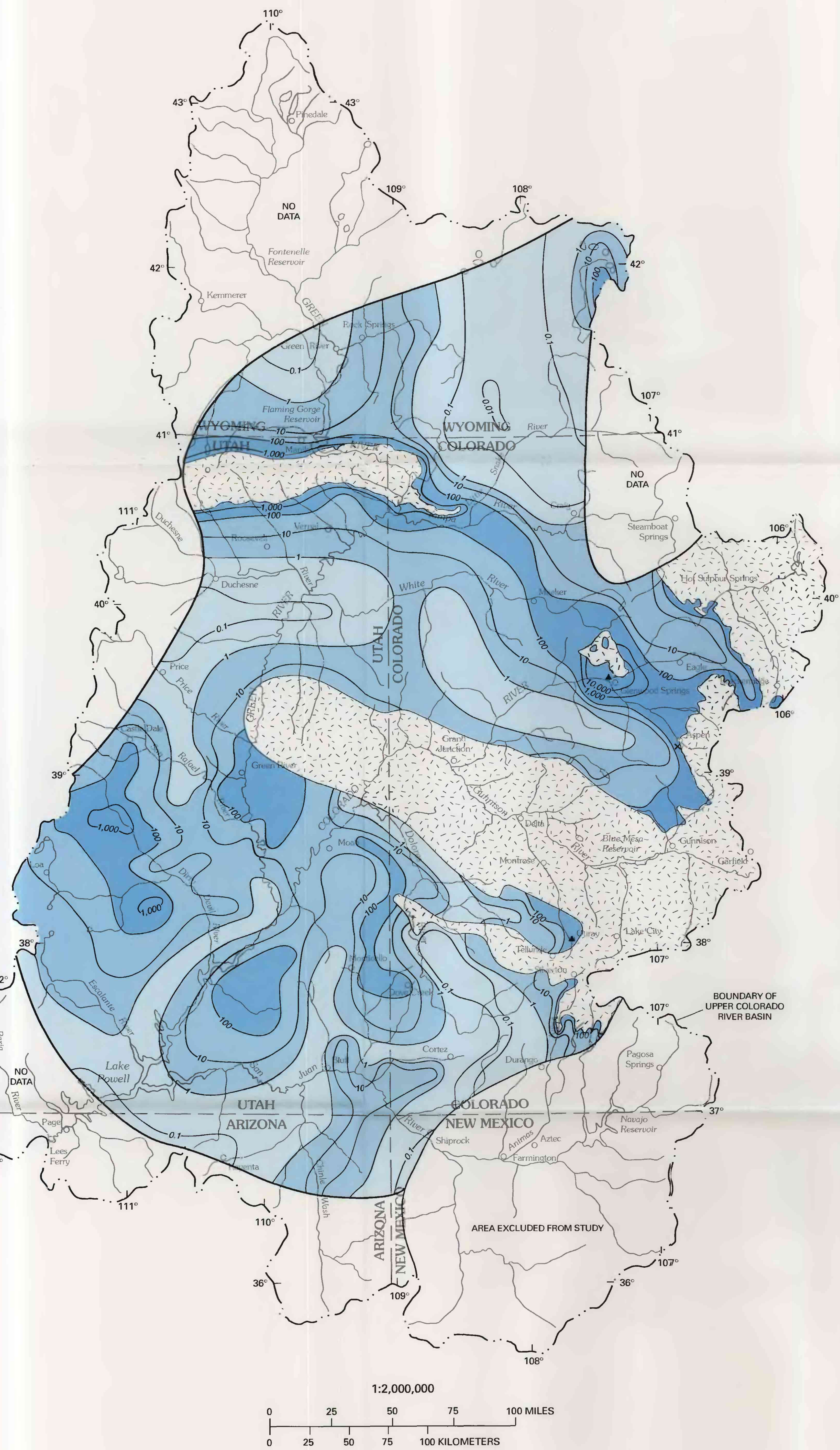
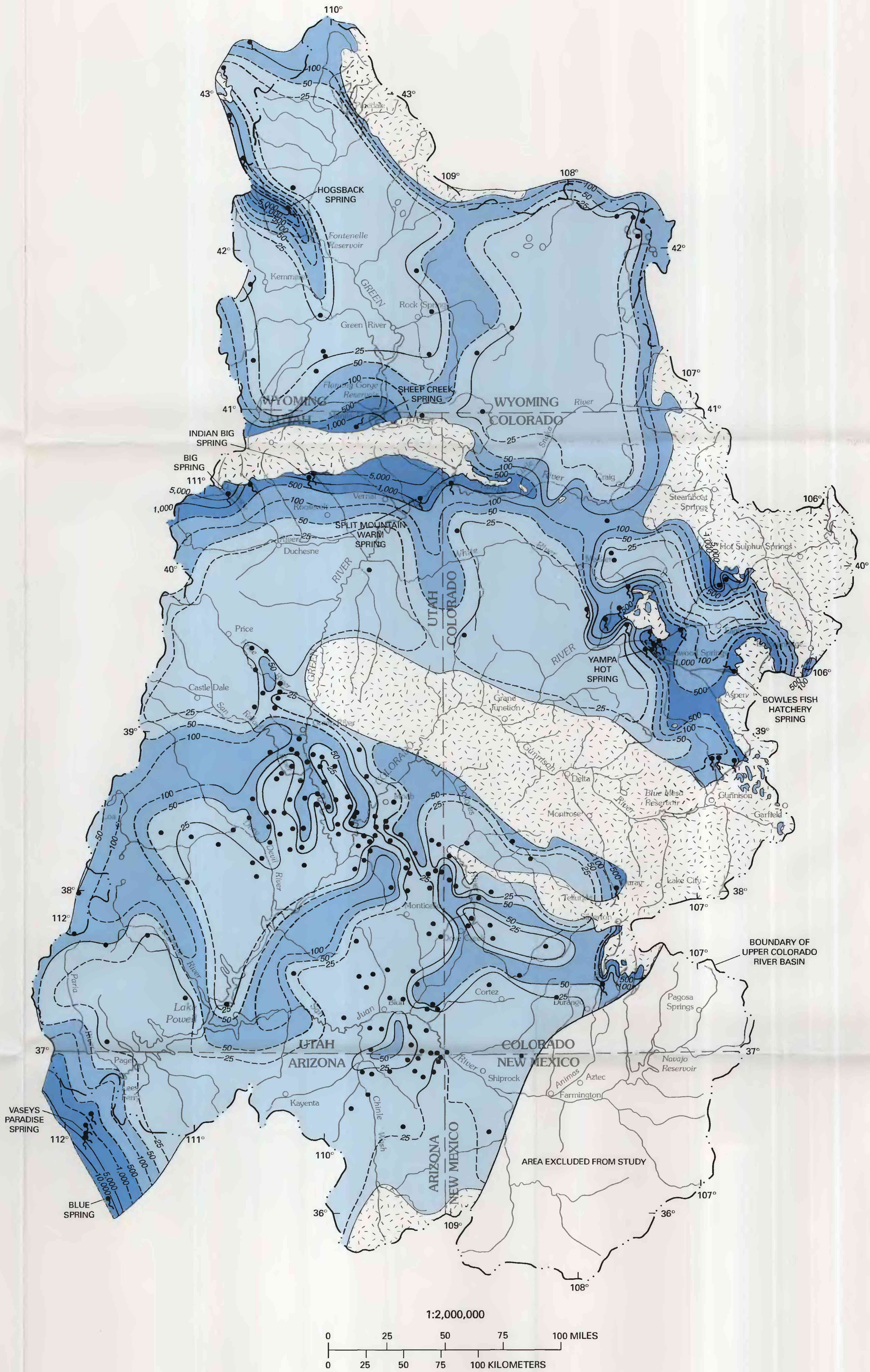


COMPOSITE TRANSMISSIVITY



MAXIMUM YIELDS

EXPLANATION

Area where the Redwall-Leadville zone is missing because of erosion or nondeposition

Composite Transmissivity

Relative composite transmissivity

- Large
- Moderate
- Small

—10— **Line of equal estimated composite transmissivity**—Location is approximate. Interval, in feet squared per day, is variable

Limit of data

Data sites

▲ Site of pumping or flowing well test from which transmissivity was calculated by analysis of drawdown- or recovery-versus-time data

× Site of pumping test with specific capacity data from which transmissivity was estimated

Maximum Yields

Relative maximum yield

- Large
- Moderate
- Small

—100— **Line of equal maximum yield from wells and springs under artesian conditions**—Dashed where approximately located. Interval, in gallons per minute, is variable

Data sites

• Borehole—Yield is flow into or from well

Spring—Labeled if discharge equals or exceeds 1,000 gallons per minute. Largest discharge (Blue Spring) is 45,000 gallons per minute

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

COMPOSITE TRANSMISSIVITY OF AND MAXIMUM YIELDS FROM THE REDWALL-LEADVILLE ZONE OF THE MADISON AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

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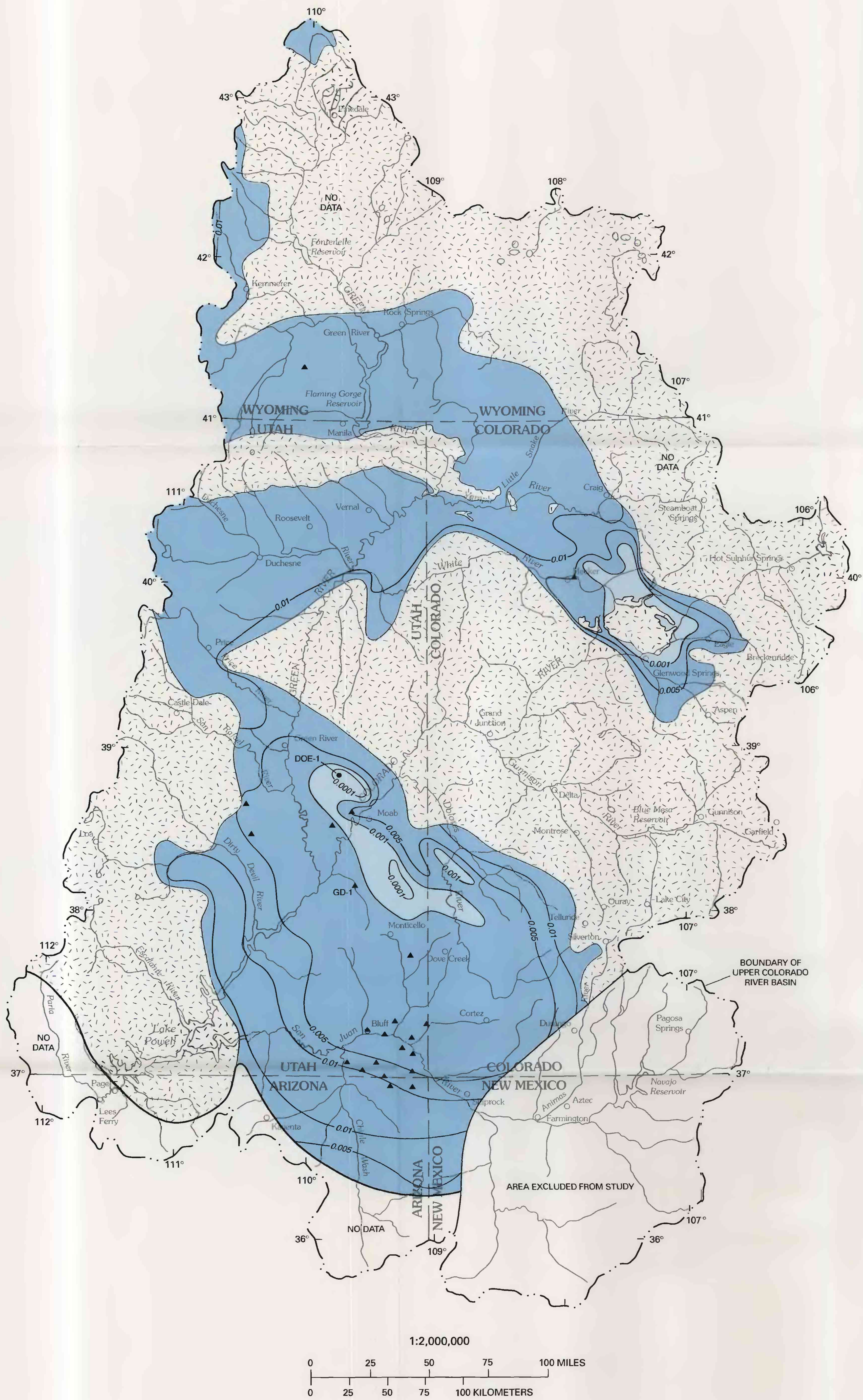
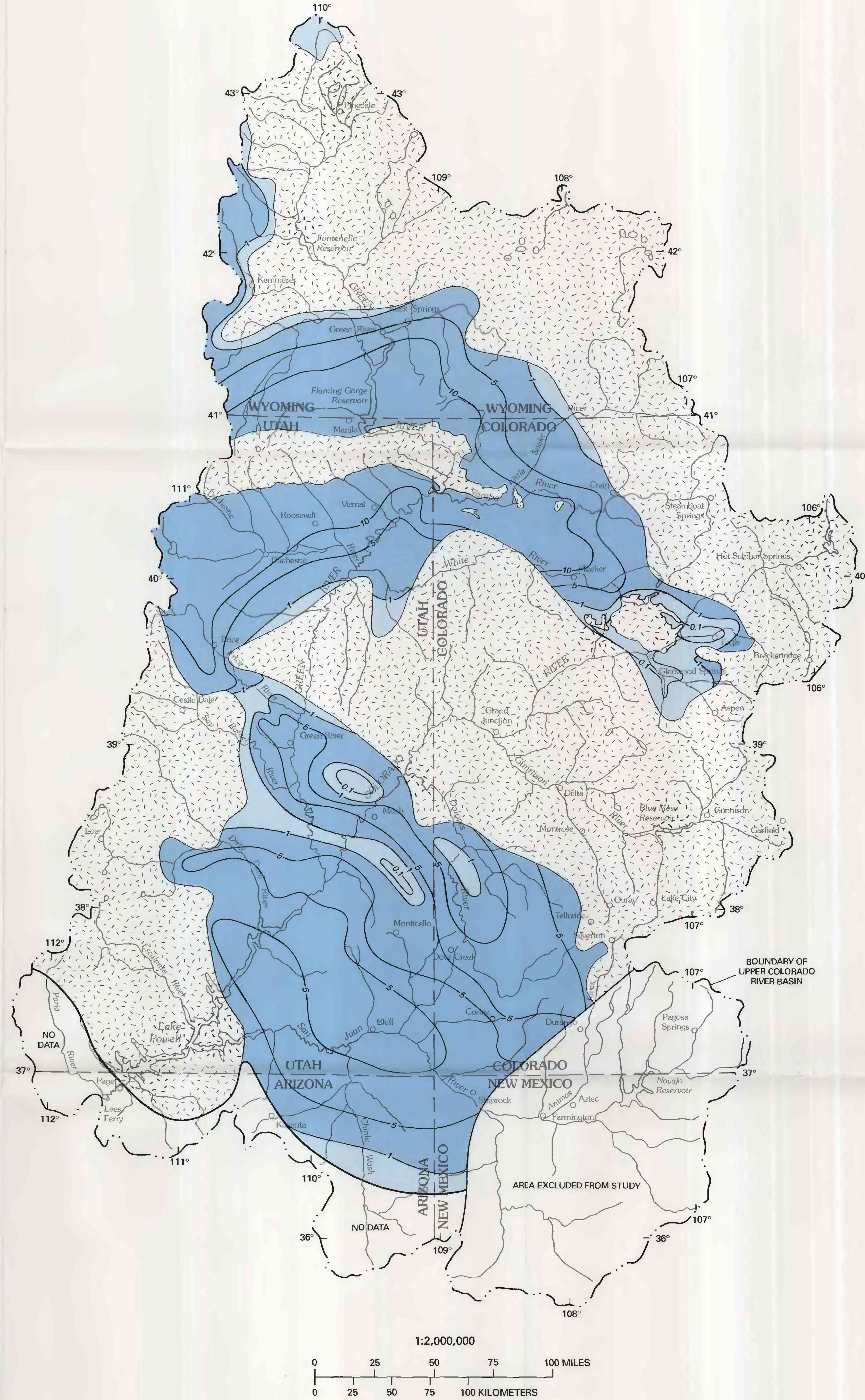


UNIT-AVERAGED HYDRAULIC CONDUCTIVITY



COMPOSITE TRANSMISSIVITY

EXPLANATION

Area where the Paradox-Eagle Valley subunit is missing because of erosion or nondeposition

Unit-Averaged Hydraulic Conductivity

Relative unit-averaged hydraulic conductivity

Moderate

Small

—0.01— Line of equal estimated unit-averaged hydraulic conductivity—Location is approximate. Interval, in feet per day, is variable

Limit of data

Data sites

• Site at which unit-averaged hydraulic conductivity was estimated from hydraulic conductivity of intervals in a borehole. Interval hydraulic-conductivity values were calculated from transmissivity values determined from slug tests

▲ Site at which unit-averaged hydraulic conductivity was estimated from permeability values obtained from drill-stem tests or laboratory measurements for borehole intervals representative of the Paradox–Eagle Valley subunit where the data were obtained

Composite Transmissivity

Relative composite transmissivity

Moderate

Small

—10— Line of equal estimated composite transmissivity—Location is approximate. Interval, in feet squared per day, is variable

Limit of data

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

UNIT-AVERAGED HYDRAULIC CONDUCTIVITY AND COMPOSITE TRANSMISSIVITY OF THE PARADOX–EAGLE VALLEY SUBUNIT OF THE FOUR CORNERS CONFINING UNIT IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

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PROFESSIONAL PAPER 1411–B
Porosity and hydraulic conductivity of the Cutler-Maroon zone—PLATE 5
Geldon, A.L., 2002, Hydrologic properties and ground-water flow systems of the Paleozoic rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, excluding the San Juan Basin

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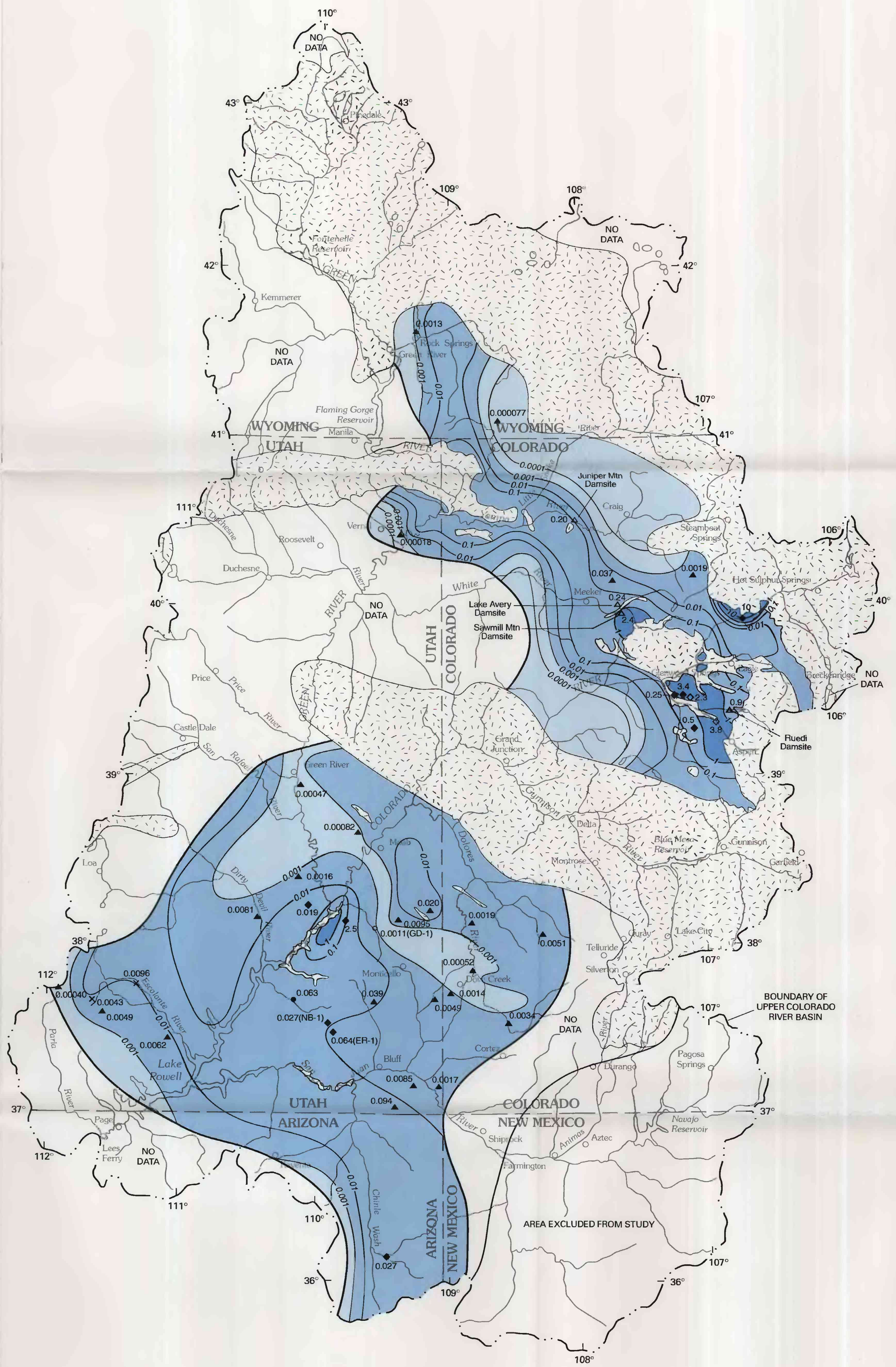
0 25 50 75 100 MILES

0 25 50 75 100 KILOMETERS

UNIT-AVERAGED POROSITY



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UNIT-AVERAGED HYDRAULIC CONDUCTIVITY

EXPLANATION

Area where the Cutler-Maroon zone is missing because of erosion or nondeposition

Unit-Averaged Porosity

—5— **Line of equal estimated unit-averaged porosity**—Location is approximate. Interval is 1 percent

•6.6 **Site at which unit-averaged porosity was estimated from the mean of laboratory-determined values in borehole intervals representative of the entire Cutler-Maroon zone at the site**—Number shown is the mean porosity of the interval, in percent

Unit-Averaged Hydraulic Conductivity

Relative unit-averaged hydraulic conductivity

- Large
- Moderate
- Small

—0.01— **Line of equal estimated unit-averaged hydraulic conductivity**—Location is approximate. Interval, in feet per day, is variable

Limit of data

Data sites—Number shown is hydraulic conductivity or average hydraulic conductivity at the site, in feet per day

- ▲ Site at which estimate was based on permeability determined by one or more drill-stem tests
- ● Site at which estimate was based on the mean of laboratory-determined permeability values in a borehole interval
- × Site at which estimate was based on the median of geophysically determined porosity values in a borehole interval
- △ Site at which estimate was based on the average of hydraulic-conductivity values determined by pressure-injection tests in a borehole or boreholes
- ◆ Site at which estimate was based on the transmissivity determined from a pumping test
- ◇ Site at which estimate was based on the transmissivity determined from a bailing test
- ○ Site at which estimate was based on the transmissivity determined from an "airlift test"

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

UNIT-AVERAGED POROSITY AND HYDRAULIC CONDUCTIVITY OF THE CUTLER-MAROON ZONE OF THE CANYONLANDS AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

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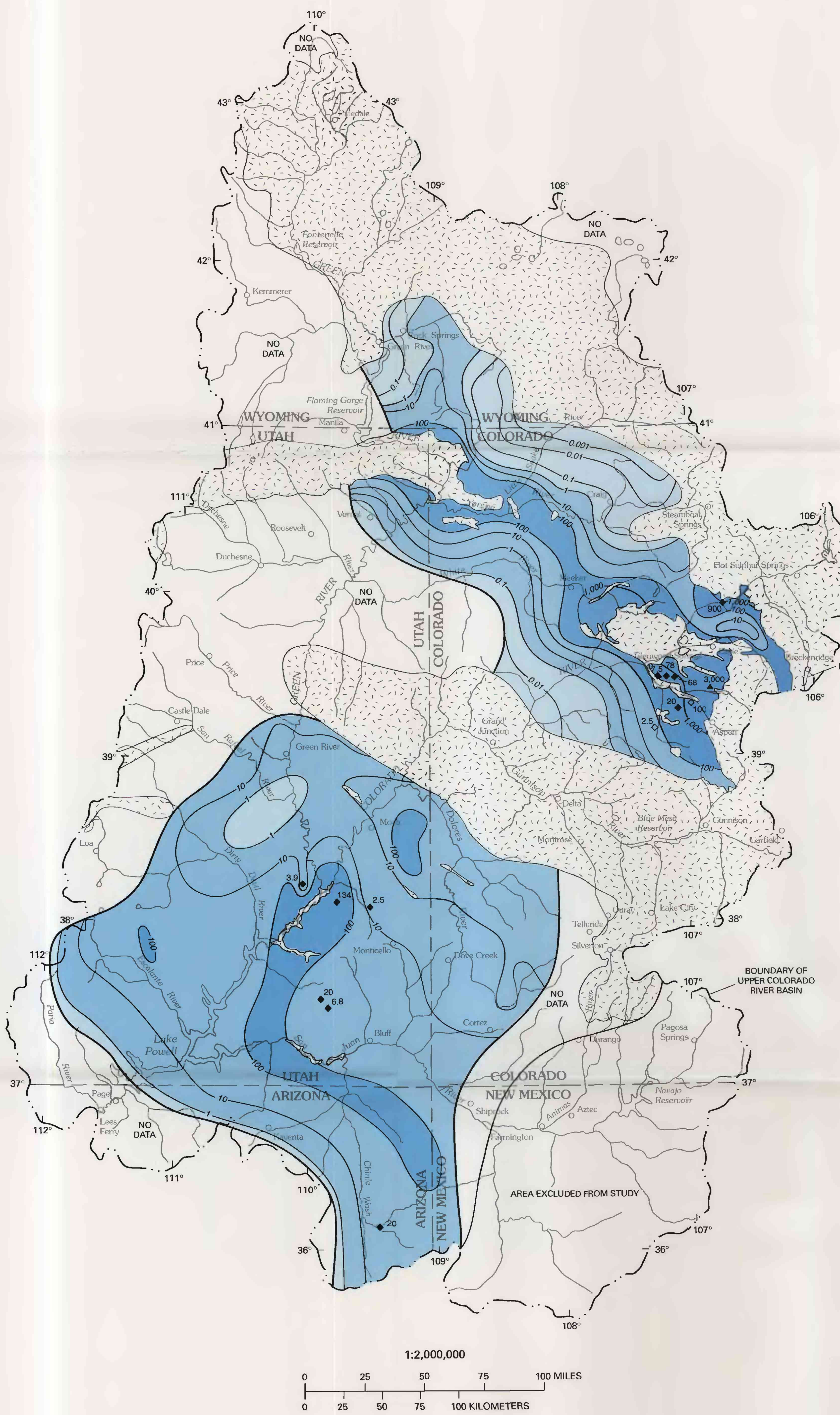
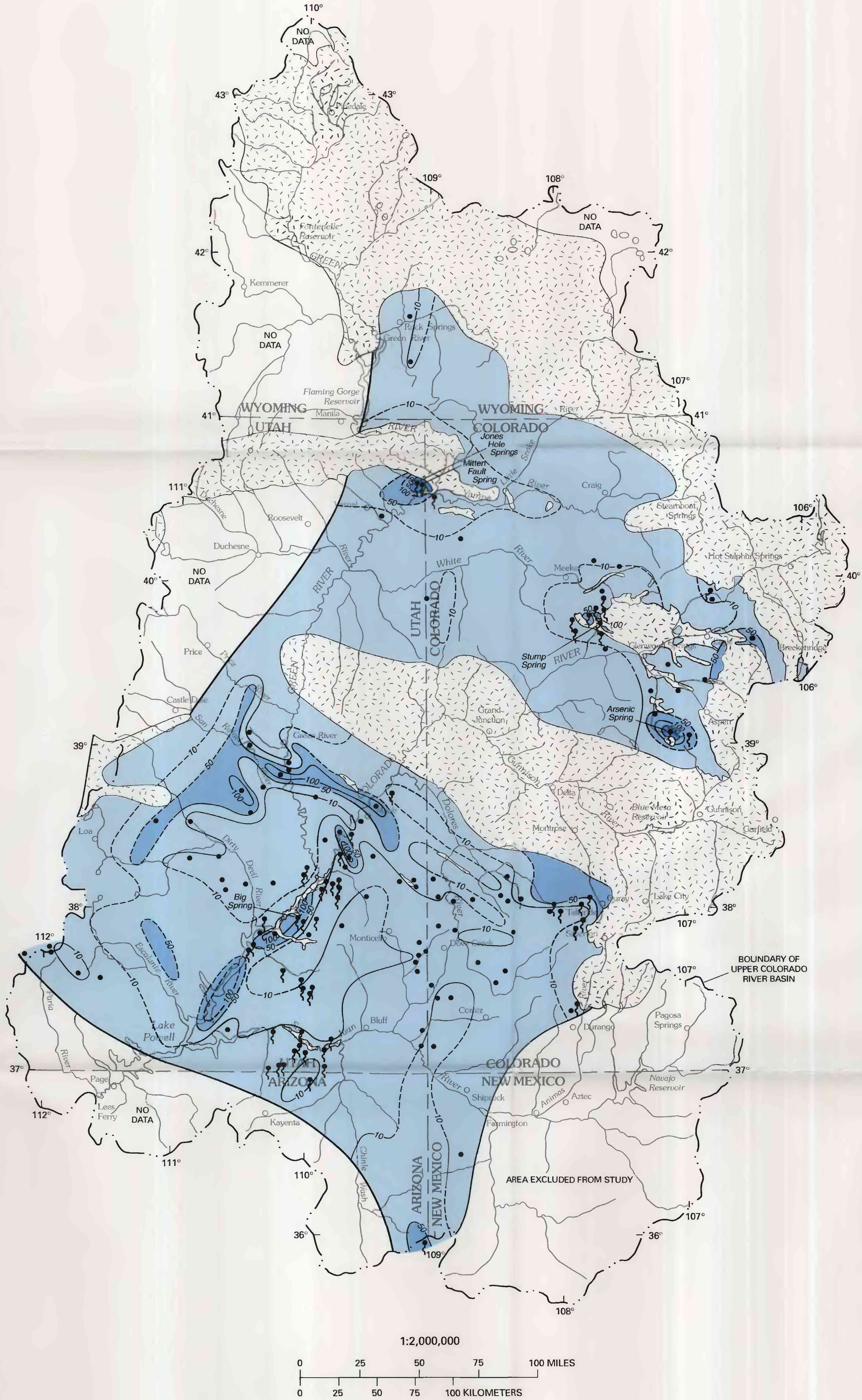


COMPOSITE TRANSMISSIVITY



MAXIMUM YIELDS

EXPLANATION

Area where the Cutler-Maroon zone is missing because of erosion or nondeposition

Composite Transmissivity

Relative composite transmissivity

- Large
- Moderate
- Small

—10— Line of equal estimated composite transmissivity—Location is approximate. Interval, in feet squared per day, is variable

Limit of data

Data sites—Number shown is transmissivity, in feet squared per day

- ◆ Site at which transmissivity was determined from a pumping test
- ◇ Site at which transmissivity was determined from a bailing test
- ▲ Site at which transmissivity was estimated from pressure injection tests and thickness determined from a cross section

Maximum Yields

Relative maximum yield

- Large
- Moderate
- Small

— 100 — — Line of equal maximum yield from wells and springs under artesian conditions—Dashed where approximately located. Interval, in gallons per minute, is variable

Limit of data

Data sites

- • Well—Discharge during a drill-stem test, flow to a well during drilling, or flow from a well after completion is indicated
- Spring—Discharge is an average at some sites. The largest springs mentioned in the text are shown

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

COMPOSITE TRANSMISSIVITY OF AND MAXIMUM YIELDS FROM THE CUTLER-MAROON ZONE OF THE CANYONLANDS AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

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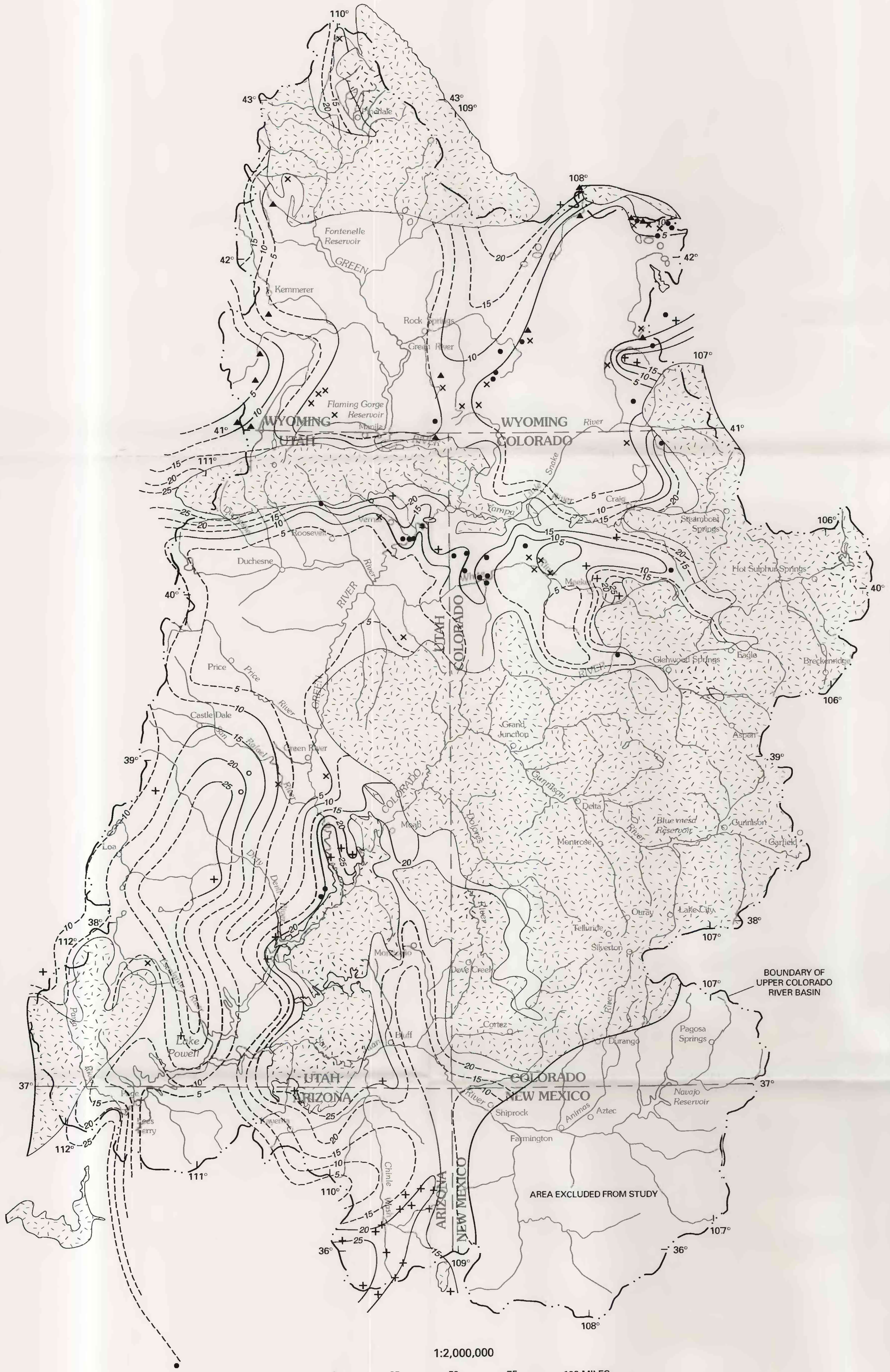
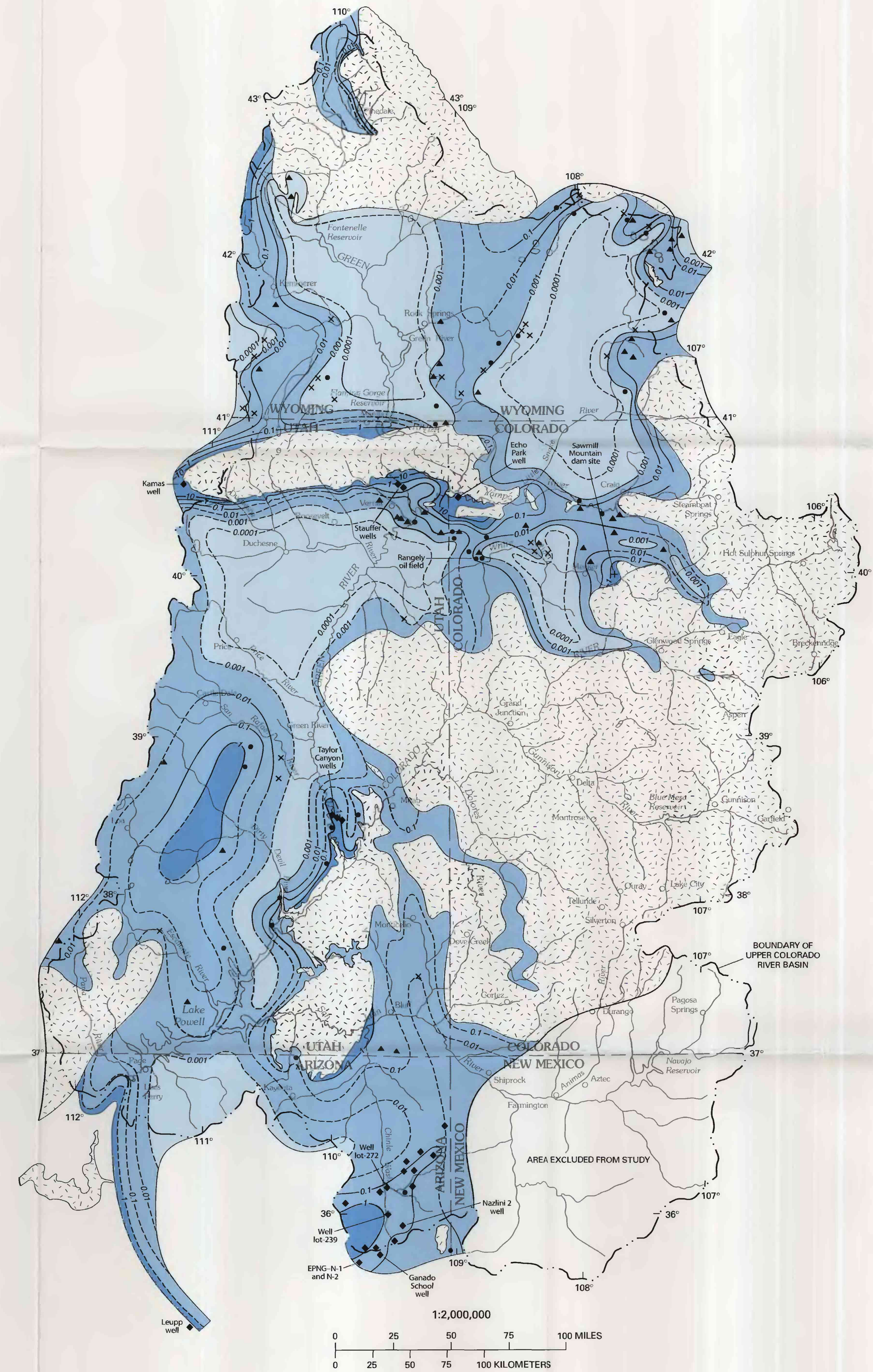


UNIT-AVERAGED POROSITY



UNIT-AVERAGED HYDRAULIC CONDUCTIVITY

EXPLANATION

Area where the Weber–De Chelly zone is missing because of erosion or nondeposition

Unit-Averaged Porosity

—5--- **Line of equal estimated unit-averaged porosity**—Dashed where approximately located. Interval is 5 percent

Data Sites

- ● Site at which esimate was based on mean porosity of samples from a borehole interval
- ○ Site at which esitmate was based on mean porosity of samples from an outcrop
- × Site at which estimate was based on median porosity in a borehole interval determined by a combination of sonic, neutron, and density logs
- ▲ Site at which estimate was based on mean porosity in a borehole interval determined by a sonic log
- + Site at which estimate was based on permeability or hydraulic conductivity using equations 7 and 11 and equations in table 3

Unit-Averaged Hydraulic Conductivity

Relative unit-averaged hydraulic conductivity

- Large
- Moderate
- Small

—1--- **Line of equal estimated unit-averaged hydraulic conductivity**—Dashed where approximately located. Interval, in feet per day, is variable

Data Sites—Labeled sites are discussed in the text or listed in table 15

- × Site at which estimate was based on median porosity in a borehole interval, as determined from geophysical logs
- ● Site at which esitmate was based on the mean laboratory-determined permeability of samples from an outcrop or borehole interval
- ▲ Site at which estimate was based on permeability determined by one or more drill-stem tests
- + Site at which estimate was based on the average of pressure-injection tests in a borehole interval
- ◆ Site at which estimate was based on transmissivity determined by a pumping, flowing-well, or bailing test

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

UNIT-AVERAGED POROSITY AND HYDRAULIC CONDUCTIVITY OF THE WEBER–DE CHELLY ZONE OF THE CANYONLANDS AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

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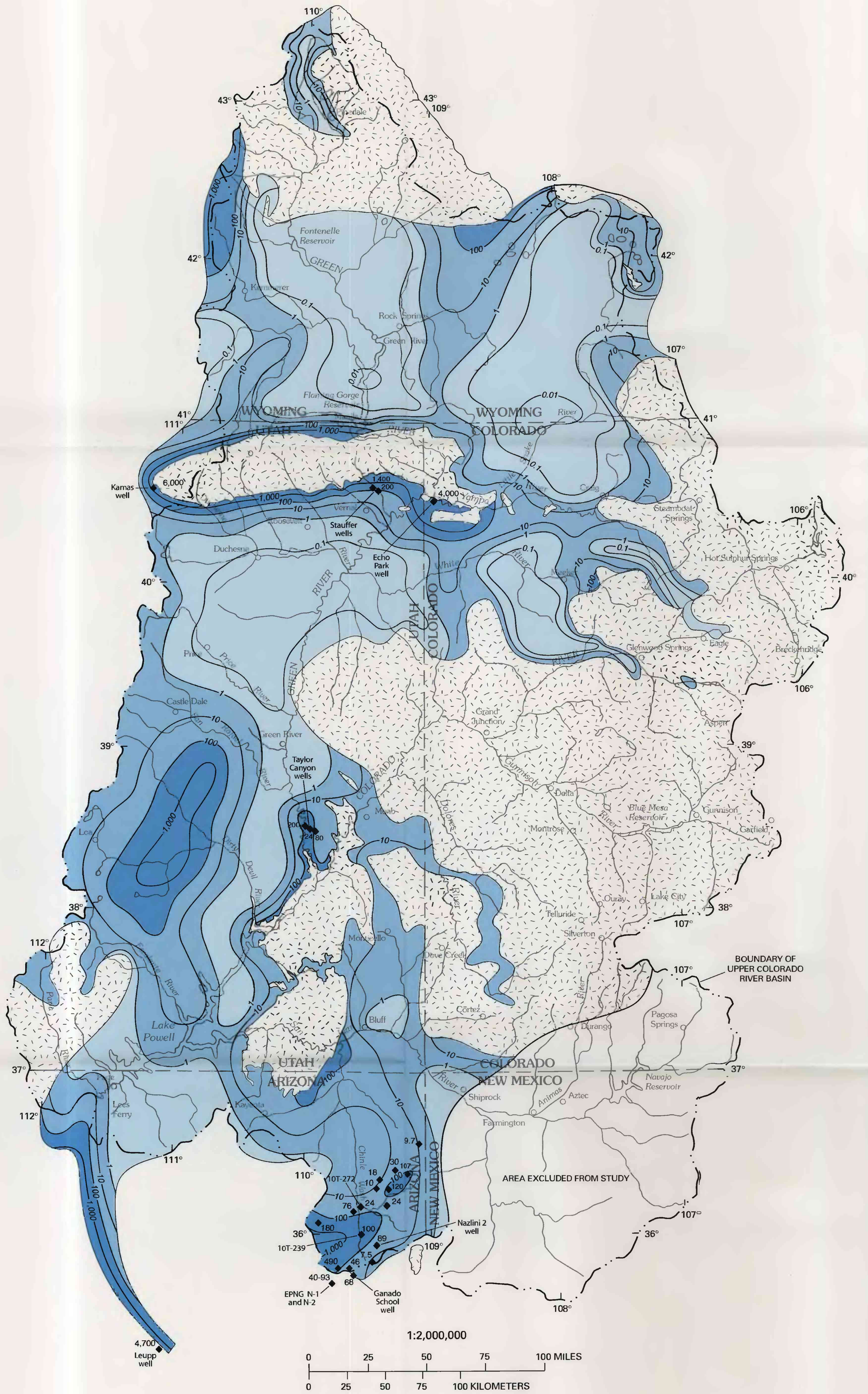
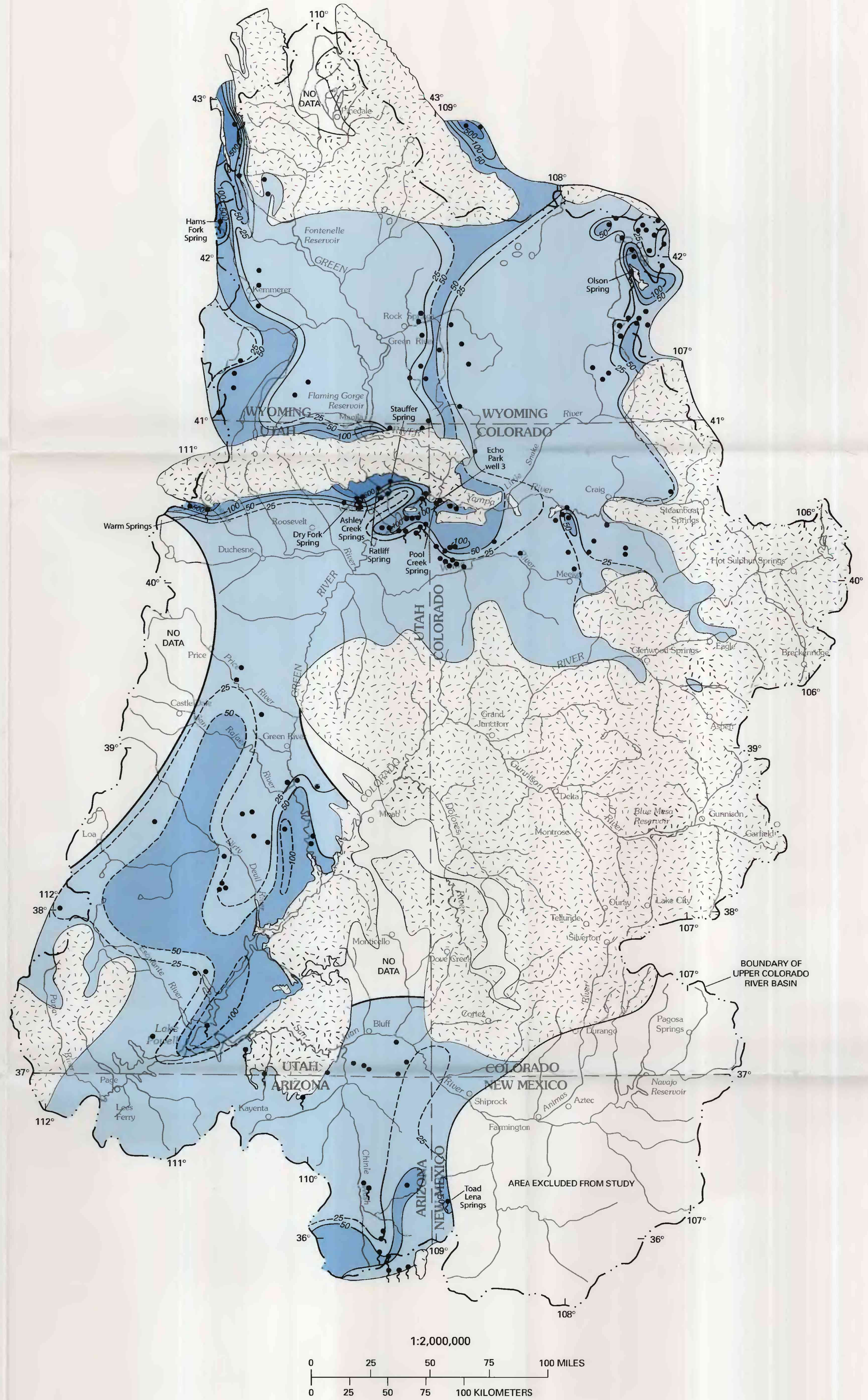
PROFESSIONAL PAPER 1411–B
Transmissivity of and maximum yields from the Weber–De Chelly zone—PLATE 8
Geldon, A.L., 2002, Hydrologic properties and ground-water flow systems of the Paleozoic rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, excluding the San Juan Basin



COMPOSITE TRANSMISSIVITY



MAXIMUM YIELDS

EXPLANATION

Area where the Weber–De Chelly zone is missing because of erosion or nondeposition

Composite Transmissivity

Relative composite transmissivity

- Large
- Moderate
- Small

—10— **Line of equal estimated composite transmissivity**—Location is approximate. Interval, in feet squared per day, is variable

6,000 **Aquifer test from which transmissivity was determined**—Number shown is transmissivity, in feet squared per day. Labeled wells are listed in table 15

Maximum Yields

Relative maximum yield

- Large
- Moderate
- Small

— 100 – – **Line of equal yield from wells and springs under artesian conditions**—Dashed where approximately located. Interval, in gallons per minute, is variable

Limit of data

Data sites

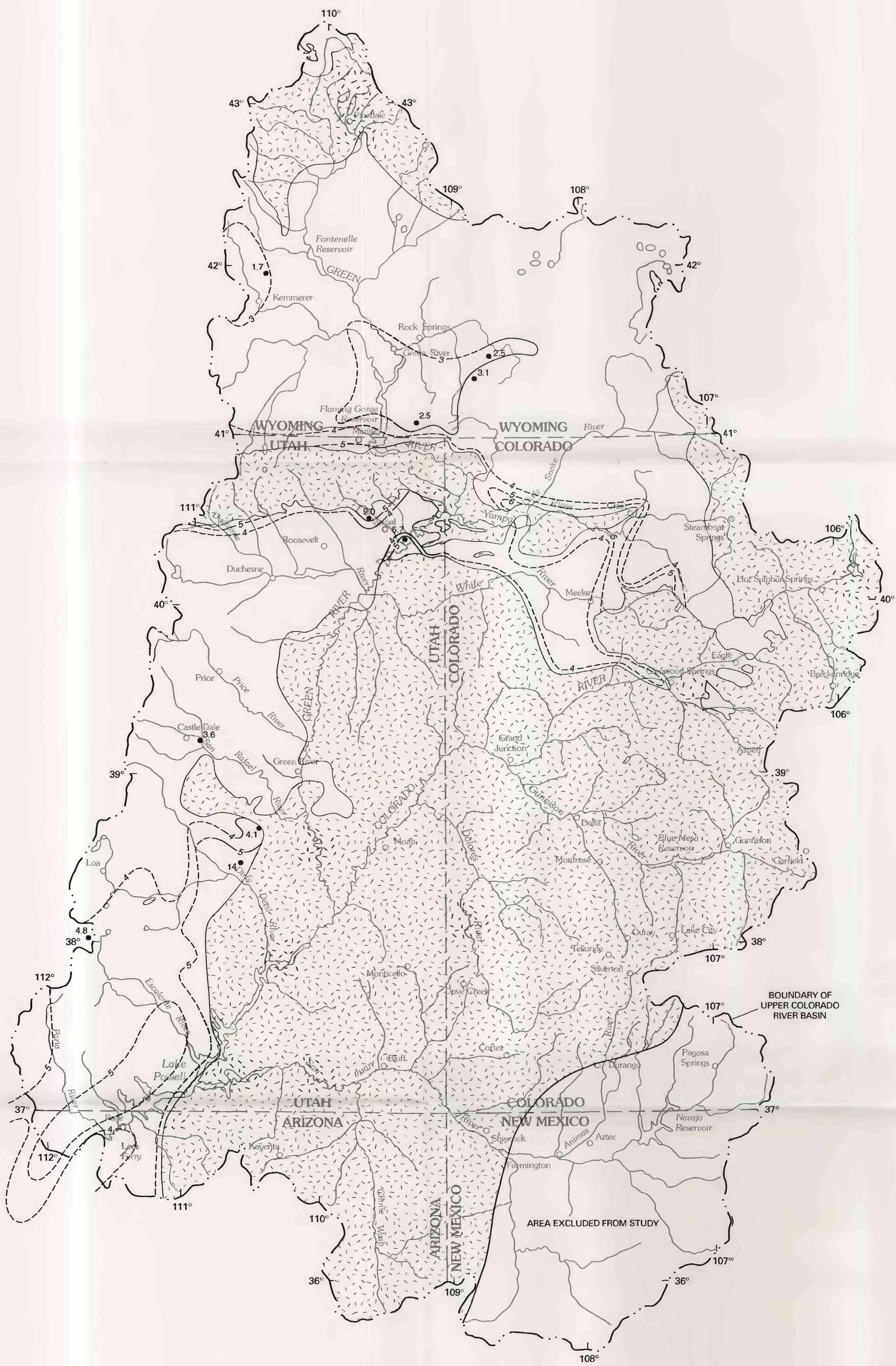
- Borehole—Discharge during a drill-stem test, flow into a well during drilling, or flow from a well after completion is indicated. Wells mentioned in text are labeled
- Spring—Moderate to large springs mentioned in the text and other large springs are labeled

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

COMPOSITE TRANSMISSIVITY OF AND MAXIMUM YIELDS FROM THE WEBER–DE CHELLY ZONE OF THE CANYONLANDS AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

By
Arthur L. Geldon
2002

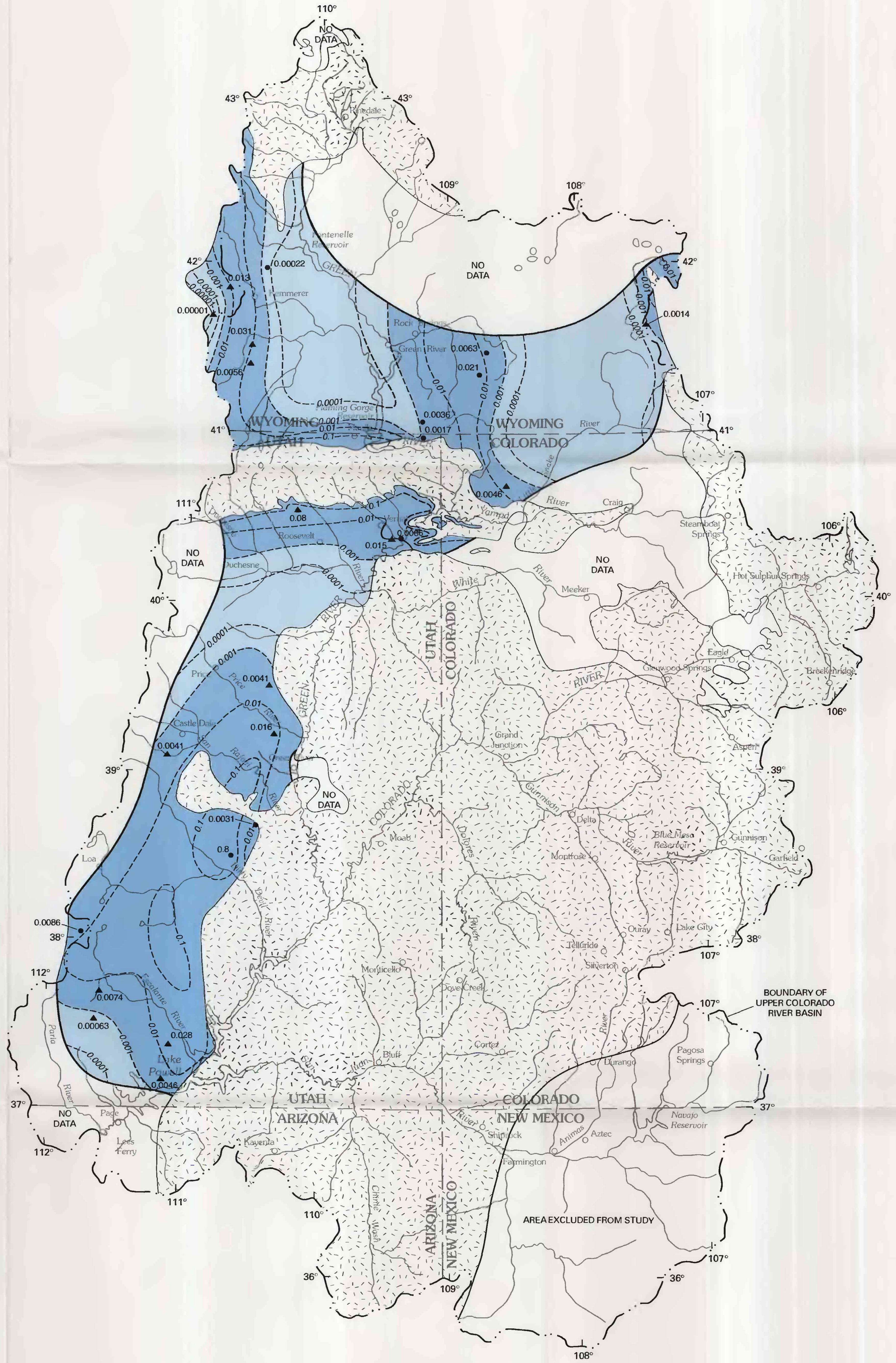
PROFESSIONAL PAPER 1411–B
Porosity and hydraulic conductivity of the Park City–State Bridge zone—PLATE 9
Geldon, A.L., 2002, Hydrologic properties and ground-water flow systems of the Paleozoic rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, excluding the San Juan Basin



1:2,000,000

0 25 50 75 100 MILES
0 25 50 75 100 KILOMETERS

UNIT-AVERAGED POROSITY



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0 25 50 75 100 MILES
0 25 50 75 100 KILOMETERS

UNIT-AVERAGED HYDRAULIC CONDUCTIVITY

EXPLANATION

Area where the Park City–State Bridge zone is missing because of erosion or nondeposition

Unit-Averaged Porosity

—3— — — **Line of equal estimated unit-averaged porosity**—Dashed where approximately located. Interval is 1 percent

3.6 **Site at which unit-averaged porosity was estimated as the mean of laboratory determined values in a borehole interval representative of the entire Park City–State Bridge zone at the site**—Number shown is the mean porosity in the interval, in percent

Unit-Averaged Hydraulic Conductivity

Relative unit-averaged hydraulic conductivity

Moderate

Small

—0.01— — **Line of equal estimated unit-averaged hydraulic conductivity**—Dashed where approximately located. Interval, in feet per day, is variable

Limit of data

Data Sites—Number shown is hydraulic conductivity, in feet per day

0.015 Site at which estimate was based on permeability determined by a drill-stem test

0.021 Site at which esitmate was based on the mean of laboratory-determined permeability values in a borehole interval

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

UNIT-AVERAGED POROSITY AND HYDRAULIC CONDUCTIVITY OF THE PARK CITY–STATE BRIDGE ZONE OF THE CANYONLANDS AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, UTAH, AND WYOMING

By
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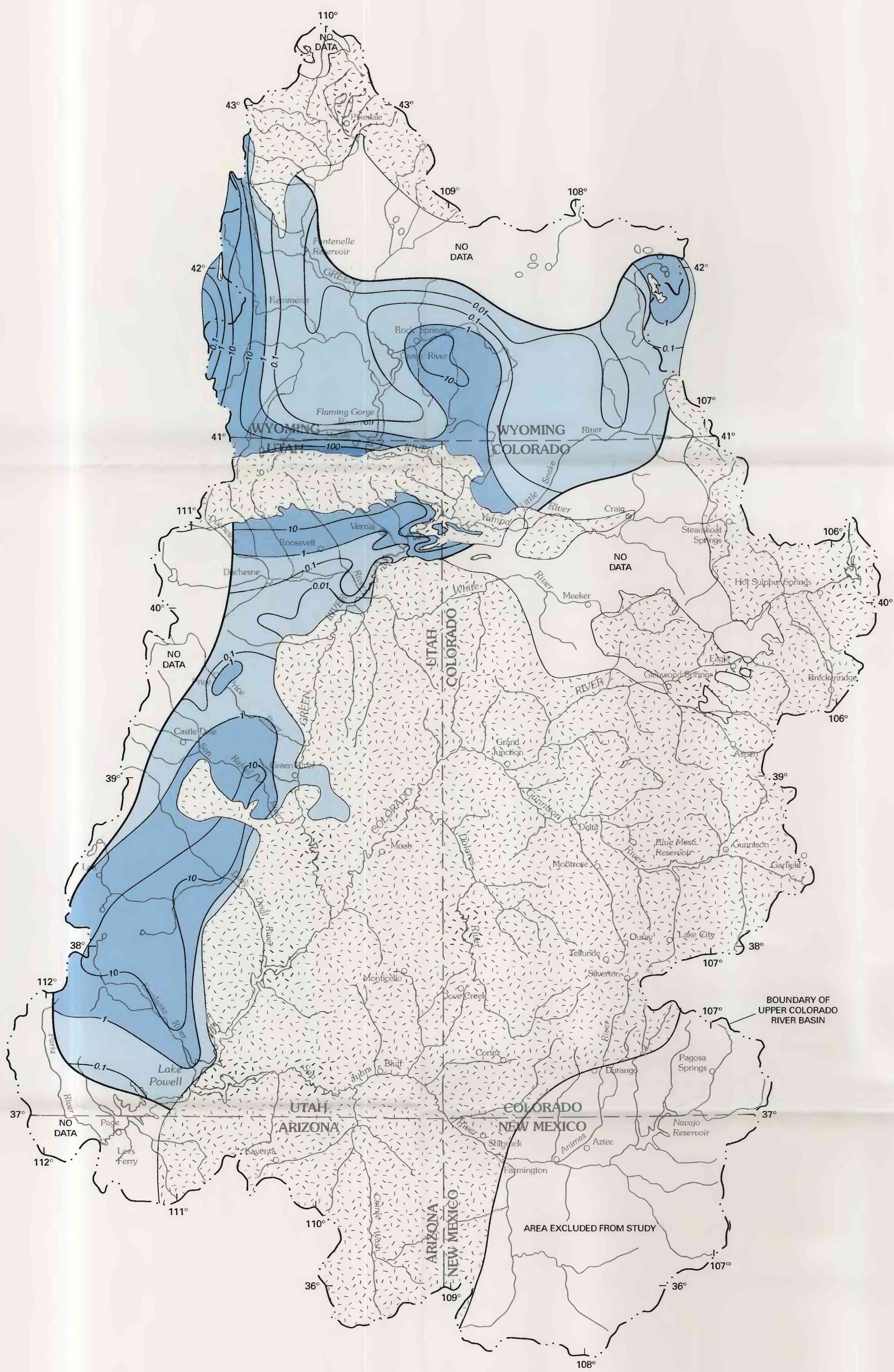
2002

U.S. DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY

PROFESSIONAL PAPER 1411–B
Transmissivity of and maximum yields from the Park City–State Bridge zone—PLATE 10
Geldon, A.L., 2002, Hydrologic properties and ground-water flow systems of the Paleozoic rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, excluding the San Juan Basin



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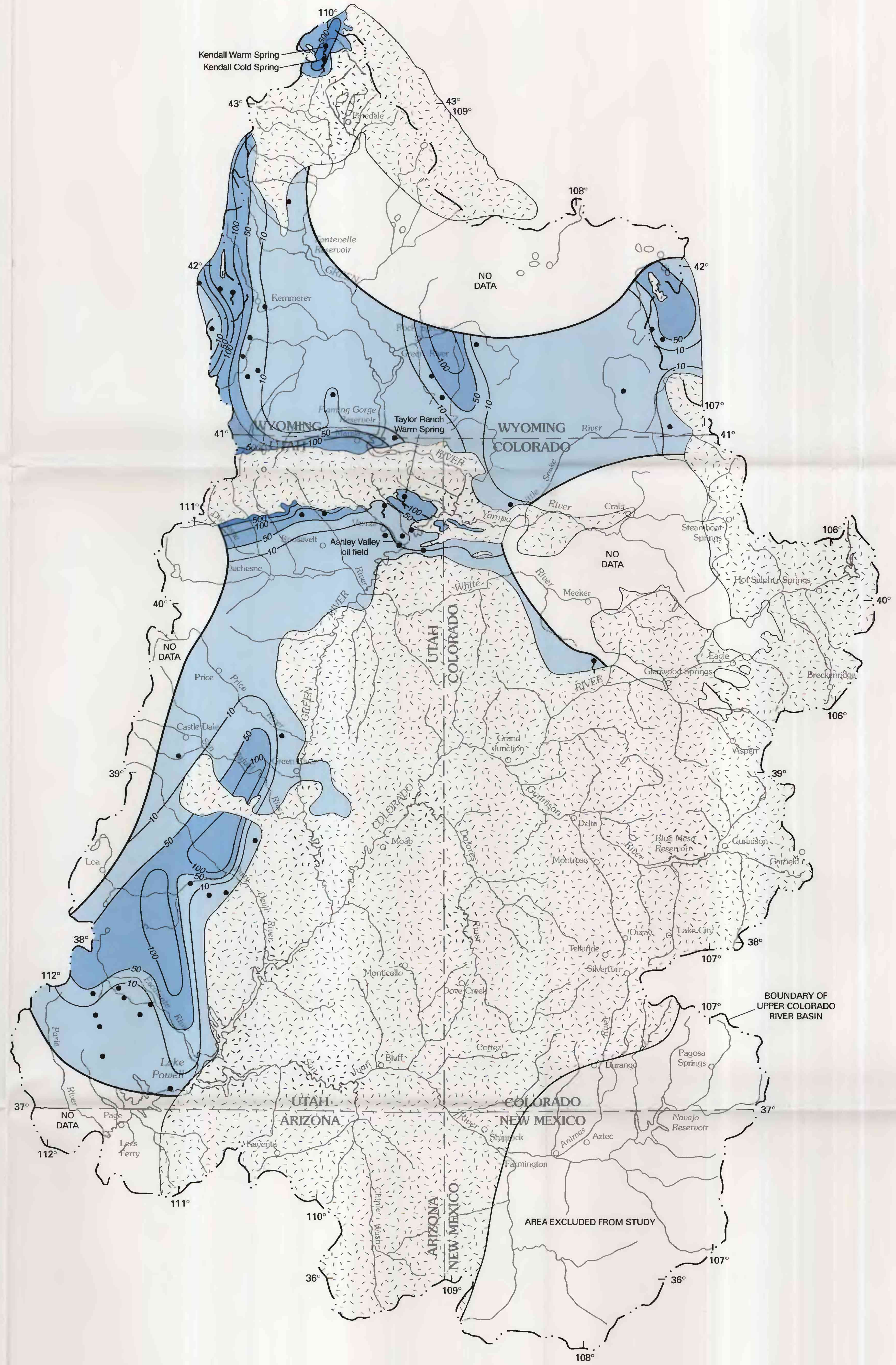
0 25 50 75 100 MILES

0 25 50 75 100 KILOMETERS

COMPOSITE TRANSMISSIVITY



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0 25 50 75 100 MILES

0 25 50 75 100 KILOMETERS

MAXIMUM YIELDS

EXPLANATION

Area where the Park City–State Bridge zone is missing because of erosion or nondeposition

Composite Transmissivity

Relative composite transmissivity

- Large
- Moderate
- Small

—10— **Line of equal estimated composite transmissivity**—Location is approximate. Interval, in feet squared per day, is variable

Limit of data

Maximum Yields

Relative maximum yield

- Large
- Moderate
- Small

—10— **Line of equal yield from wells and springs under artesian conditions**—Location is approximate. Interval, in gallons per minute, is variable. Map is based on maximum measured, reported, or estimated values in boreholes or survey sections

Limit of data

Data sites

• Borehole—Discharge during a drill-stem test, flow into a well during drilling, or flow from a well without drawdown after completion is indicated

Spring—Large springs mentioned in the text are labeled

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

COMPOSITE TRANSMISSIVITY OF AND MAXIMUM YIELDS FROM THE PARK CITY–STATE BRIDGE ZONE OF THE CANYONLANDS AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, UTAH, AND WYOMING

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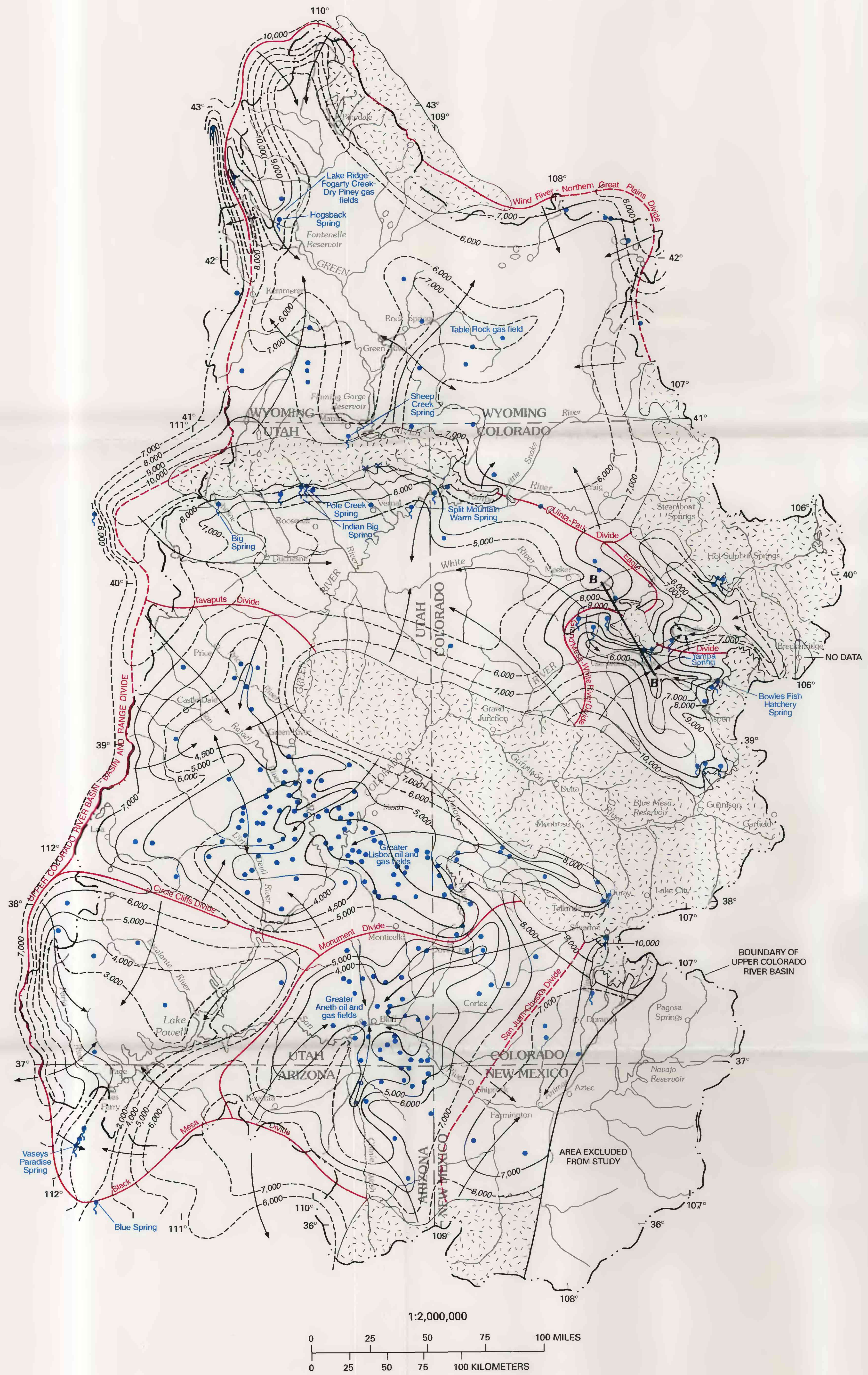
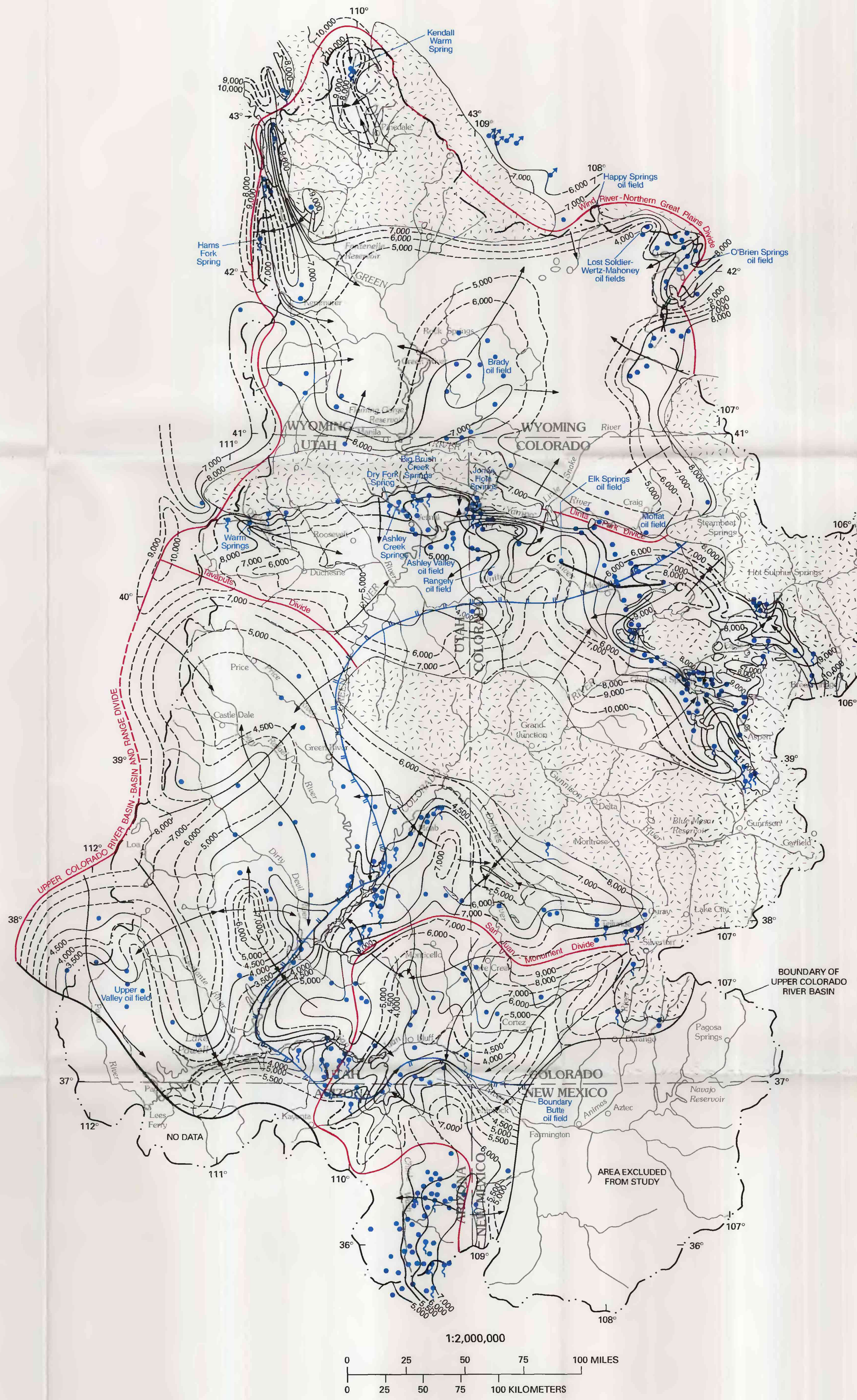


REDWALL-LEADVILLE ZONE OF MADISON AQUIFER



WEBER–DE CHELLY AND CUTLER-MAROON ZONES OF CANYONLANDS AQUIFER

EXPLANATION

Area where the Redwall-Leadville, Weber–De Chelly, or Cutler-Maroon zone is missing because of erosion or nondeposition

5,000 **Potentiometric contour**—Shows altitude at which water level would have stood in tightly cased wells. Dashed where approximately located. Arrow indicates direction of ground-water movement assuming the aquifer is isotropic. Contour interval, in feet, is variable. National Geodetic Vertical Datum of 1929. For the Weber–De Chelly and Cutler-Maroon aquifers, potentiometric contours south of latitude 37° are based in part on Cooley and others (1969, fig. 9); Levings and Farrar (1977a, 1977b, 1977c); and Eychaner (1983, fig. 3); potentiometric contours between Grand Junction and Glenwood Springs, Colorado, are based in part on topography

Limit of data

Ground-water divide—Dashed where approximately located

B —— *B'* **Location of hydrogeologic section shown in figures 94 and 98**

Eastern extent of Weber–De Chelly zone as a fully saturated, productive hydrogeologic unit—East of this line the Weber–De Chelly zone pinches out and where present is partially saturated to unsaturated. The principal productive zone in the Canyonlands aquifer east of the line is the Cutler-Maroon zone. Heads shown west of the line are for the Weber–De Chelly zone. Heads shown east of the line are for the Cutler-Maroon zone. At the line heads in the Weber–De Chelly and Cutler-Maroon zones are equal, and heads for both zones are shown

Data sites

Borehole—Static water level is plotted

Flowing well—Altitude of land surface is plotted if static water level is not known

Spring—Altitude of land surface is plotted. Springs with discharges of 1,000 gallons per minute or more are plotted

× Sink—Altitude of land surface is plotted

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

POTENTIOMETRIC SURFACES OF THE MADISON AND CANYONLANDS AQUIFERS IN THE UPPER COLORADO RIVER BASIN AND VICINITY, IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

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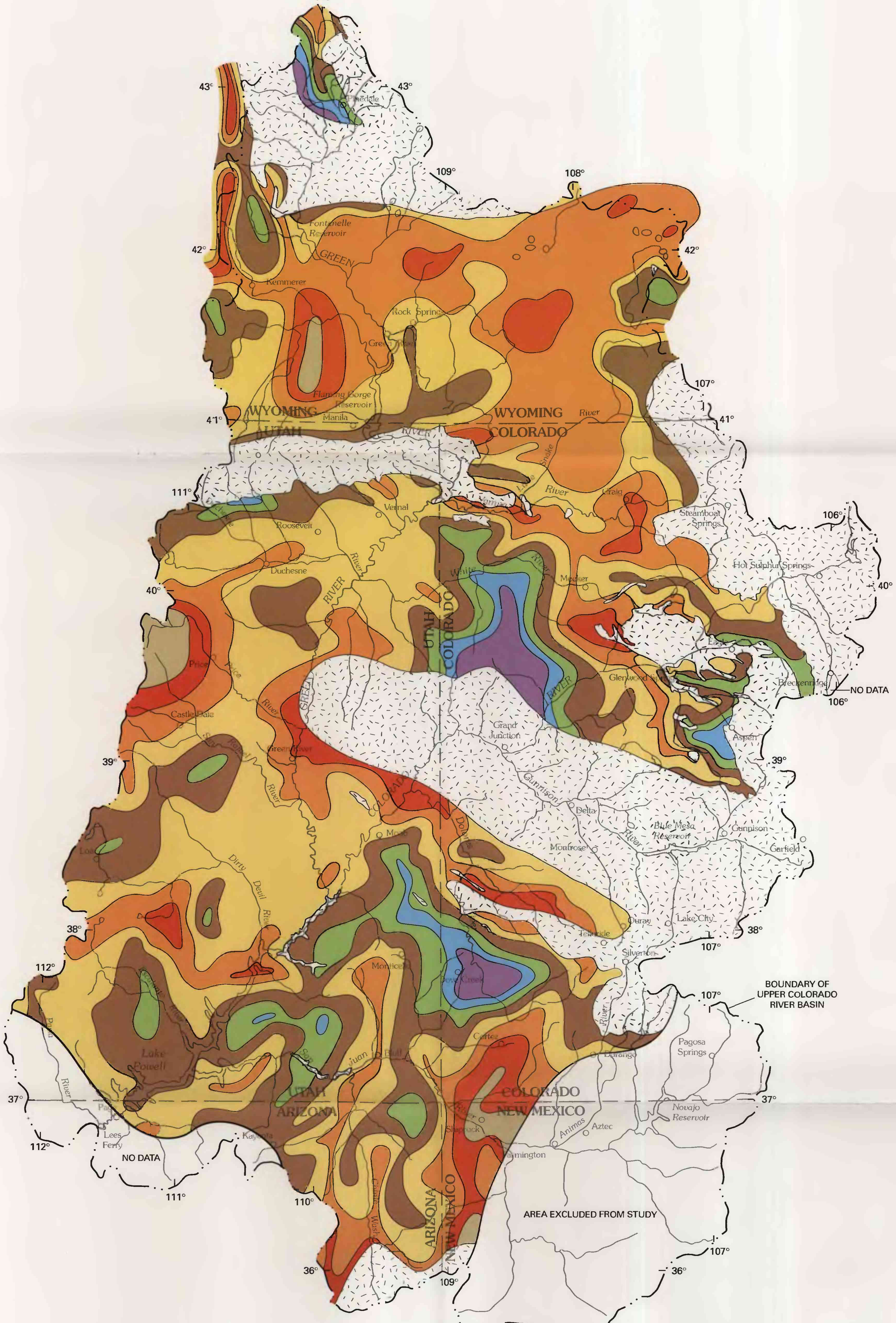
2002



EXPLANATION

Area where either the Canyonlands aquifer, Madison aquifer, or both aquifers are missing because of erosion, nondeposition, or faulting

Difference, in feet, between potentiometric heads in the Canyonlands aquifer and Four Corners aquifer system

- 3,500 to 4,500
- 2,500 to 3,500
- 1,500 to 2,500
- 500 to 1,500
- –500 to 500
- –1,500 to –500
- –2,500 to –1,500
- –3,500 to –2,500

Limit of data

1:2,000,000

0 25 50 75 100 MILES

0 25 50 75 100 KILOMETERS

Base from U.S. Geological Survey U.S. base map, 1:2,500,000

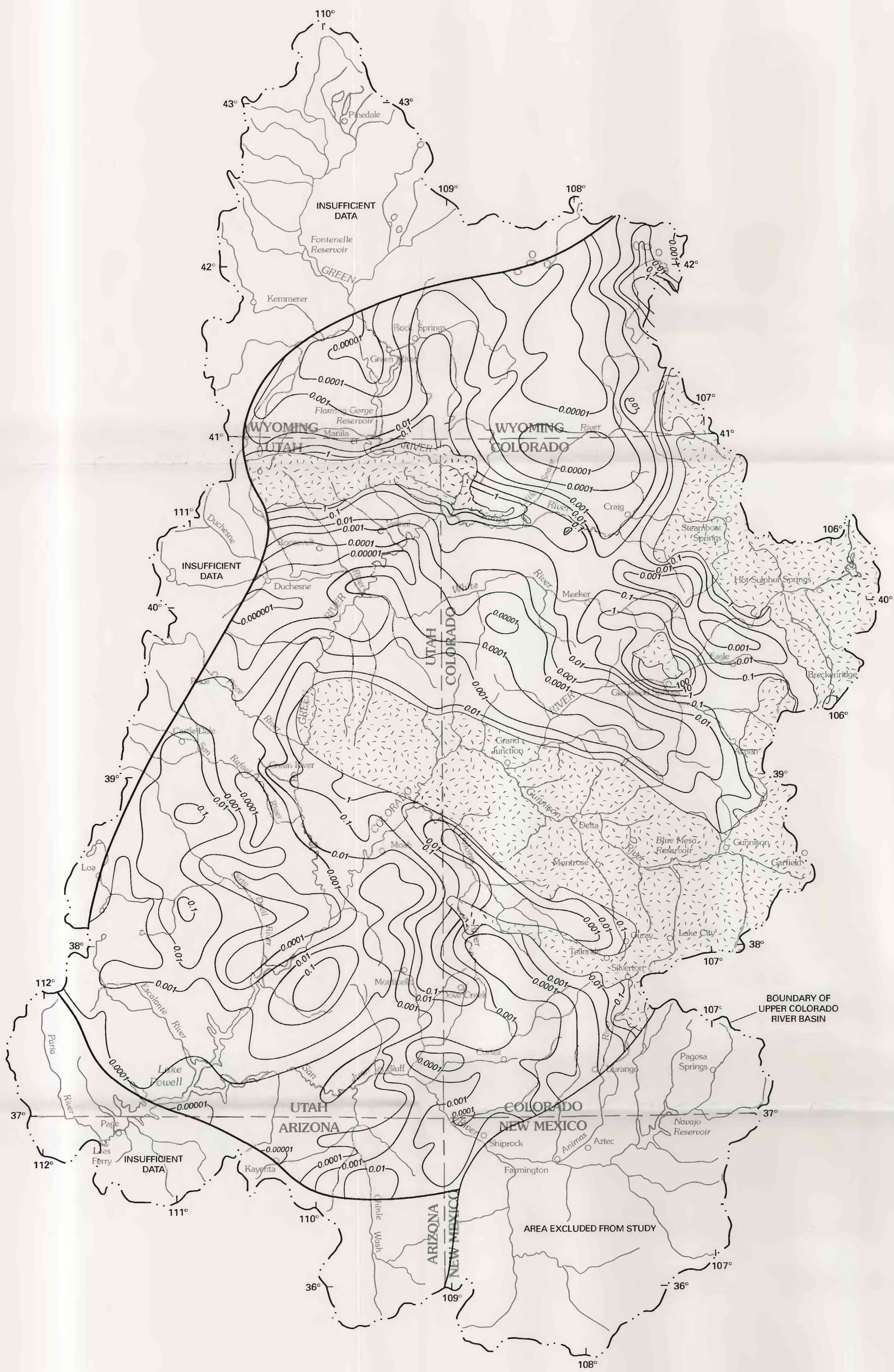
DIFFERENCE BETWEEN POTENTIOMETRIC HEADS IN THE CANYONLANDS AQUIFER AND MADISON AQUIFER IN THE UPPER COLORADO RIVER BASIN AND VICINITY IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

By

Arthur L. Geldon

2002

PROFESSIONAL PAPER 1411–B
Linear velocity of water in the Redwall-Leadville and Weber–De Chelly zones—PLATE 13
Geldon, A.L., 2002, Hydrologic properties and ground-water flow systems of the Paleozoic rocks in the Upper Colorado River Basin in Arizona, Colorado, New Mexico, Utah, and Wyoming, excluding the San Juan Basin



1:2,000,000

0 25 50 75 100 MILES

0 25 50 75 100 KILOMETERS

REDWALL-LEADVILLE ZONE OF MADISON AQUIFER

1:2,000,000

0 25 50 75 100 MILES

0 25 50 75 100 KILOMETERS

WEBER–DE CHELLY ZONE OF CANYONLANDS AQUIFER

EXPLANATION

Area where the Redwall-Leadville or Weber–De Chelly zone is missing or unsaturated

—0.001— Line of equal estimated average linear velocity—Interval, in feet per day, is variable

Limit of data

Base from U.S. Geological Survey
U.S. base map, 1:2,500,000

AVERAGE LINEAR VELOCITY OF WATER IN THE REDWALL-LEADVILLE ZONE OF THE MADISON AQUIFER AND WEBER–DE CHELLY ZONE OF THE CANYONLANDS AQUIFER IN THE UPPER COLORADO RIVER BASIN IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING

By
Arthur L. Geldon
2002